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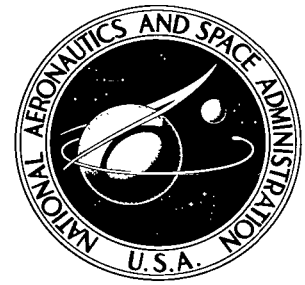

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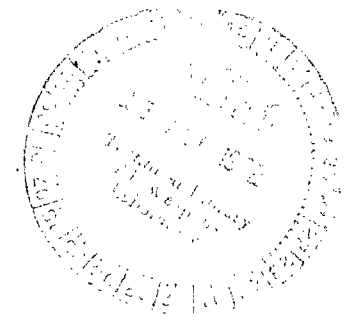
METEOROLOGICAL FLIGHT CONDITIONS FOR SUPERSONIC AIRCRAFT

by I. G. Pchelko, N. V. Petrenko, and G. S. Buldovskiy



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METEOROLOGICAL FLIGHT CONDITIONS FOR SUPERSONIC AIRCRAFT

By I. G. Pchelko, N. V. Petrenko,
and G. S. Buldovskiy

Translation of "Meteorologicheskiye Usloviya
Poletov Sverkhzvukovykh Samoletov."
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PREFACE

Current development of aviation places before meteorologists progressively more new problems, one of which pertains at the present time to meteorological flight safety of supersonic transports. Presumably it will not be long until supersonic transports will be making regularly scheduled flights in the stratosphere at altitudes of about 20 km and for distances of several thousand kilometers. In this connection, the weather service will be required to furnish complete information and forecasts pertaining to the distribution of temperature fields, winds, turbulence zones and other elements and phenomena of the weather over a considerable part of the northern hemisphere in the layer of the atmosphere up to altitudes of 20 to 25 km.

In order to satisfy the imposed requirements it will be essential to know the features of atmospheric circulation in the stratosphere during the different seasons of the year, the relationship between these processes and processes in the troposphere, the influence of individual meteorological elements and phenomena on supersonic transport flight conditions during takeoff and ascent, cruise, descent and landing.

During the last decade, according to observations during the International Geophysical Year and Year of International Cooperation, both in the USSR and in other countries, comprehensive investigations have been made of the features of atmospheric circulation in the stratosphere and its interaction with circulation in the troposphere. The results of the investigation for the possible influence of meteorological conditions and solar flares, conducted over the period 1965-1968, have been analyzed both at international and at intra-union scientific conferences and symposia.

The basic results of investigations that have been conducted and of discussions are presented in this book, which therefore should be a concise text for aviation meteorologists, pilots and other aeronautical specialists.

Chapters 1 and 2 are written by I. G. Pchelko, Chapters 3, 4 and 6 by N. V. Petrenko and Chapter 5 by G. S. Buldovskiy.

The authors express their sincere gratitude to S. V. Solonin, who devoted himself to review the preliminary manuscript of this monograph and to editing the final manuscript, and also for his valuable advice, which facilitated improvement of the content of this text.

METEOROLOGICAL FLIGHT CONDITIONS FOR SUPERSONIC AIRCRAFT

I. G. Pchelko, N. V. Petrenko and G. S. Buldovskiy

ABSTRACT: This book is concerned with atmospheric structure; emphasis is placed on certain important features of meteorological conditions in the troposphere and stratosphere. Seasonal pecularities of atmospheric circulation in the stratosphere are described. Interaction between stratospheric and atmospheric processes is analyzed. Meteorological phenomena and meteorological elements at all stages of supersonic flight are discussed. The book will be of interest to aeronautical meteorologists. It will also be useful to pilots and specialists interested in meteorological flight conditions in the stratosphere.

CHAPTER 1

/4*

THE ATMOSPHERE, ITS COMPOSITION AND STRUCTURE

The atmosphere is the medium in which aircraft fly and which has a tremendous influence on their performance characteristics. Therefore, a thorough knowledge of its composition, structure and features of circulation during various seasons of the year is essential.

The earth's surface represents, in a manner of speaking, the floor of the vast air ocean, and all life on it owes its existence to the atmosphere. It supplies the oxygen required for breathing, protects us from the lethal effects of cosmic rays and ultraviolet radiation, and also protects the earth's surface from severe heating in the day and severe cooling at night.

Without the atmosphere the temperature of the earth's surface would go up to 100° and more by day, and at night the temperature would drop sharply to -100° Total silence would prevail everywhere, since sound cannot propagate in a vacuum, day would replace night instantaneously, and the sky would be absolutely black. The property of particles of atmospheric air and various foreign impurities in the air to "break down" sunlight into a multiplicity of small rays

*Numbers in the margin indicate pagination in the foreign text.

and to scatter it in every direction is responsible for the blue of the sky, variegation of the glow, etc.

The atmosphere is transparent, but it constantly reminds us of its presence: rain, snow, thunderstorms and blizzards, hurricanes and calms, heat waves and cold snaps, all are manifestations of atmospheric processes that take place under the influence of uneven distribution of solar energy and during interaction between the atmosphere and the very surface of the earth.

The atmosphere has no distinct upper boundary; the density of its constituent gases gradually approaches that of the gases of interplanetary space. Much of the atmosphere's mass, however, is concentrated in the comparatively thin layer adjacent to the earth's surface. About 50% of its entire mass is contained in the layer from the earth's surface to an altitude of 5 km, 75%, up to 10 km, and 90% up to 16 km. The small thickness of the basic (in terms of mass) layer of the atmosphere, compared to its horizontal expanse, results in the fact that the vertical scales of the phenomena and processes that occur in the atmosphere have substantially less influence on their horizontal dimensions. Air masses, cyclones and anticyclones, also fronts have a horizontal reach of hundreds and thousands of kilometers, but extend vertically only a few kilometers. /5

1.1. Composition of the Atmosphere

The atmosphere consists of a mixture of gases, called air, in which are suspended liquid and solid particles. The total mass of these particles is slight compared to the basic mass of the atmosphere. Atmospheric air near the earth's surface is, as a rule, humid. This means that it contains water vapor along with other gases. The concentration of water vapor in the air varies in a broad range, in contrast to other components of air: at the earth's surface it varies from hundredths of a percent to several percent. The reason for this is that under existing conditions water vapor can convert to the liquid and solid states and, conversely, can reenter the atmosphere as a result of evaporation from the earth's surface. Air that does not contain water vapor is called dry air.

Dry air consists of the following gases: nitrogen (78% by volume, or 76% by mass), oxygen (21% by volume or 23% by mass), argon (0.93%), carbon dioxide (0.03%). These are followed by gases whose concentrations are negligible: neon (0.0018%), helium (0.0004%), ozone (0.000003%), and traces of krypton and xenon.

In addition to the above gaseous components, the atmosphere contains many foreign impurities: smokes and industrial gases, smoke from forest fires, vast quantities of dust of various origin, including products of radioactive decay, pollen and plant seeds, spores, fine particles of marine salt, etc.

Analyses of air samples taken as early as 1933 during the ascent of the stratosphere balloon, USSR-1 first revealed that the composition of the atmosphere at 20 km is exactly the same as its composition near the earth's surface. Numerous subsequent investigations established that the atmosphere is uniformly mixed up to considerable altitudes and that the composition of the atmosphere does not change substantially up to an altitude of about 95 km.

Ozone in the Atmosphere

Only ozone, which is a colorless (in small concentrations it is blue) gas with the characteristic sharp odor, behaves differently. Near the earth's surface the concentration of ozone is extremely low (about $2 \cdot 10^{-8}$% by volume) and is produced chiefly by the breaking down, dissociation or separation of oxygen molecules into atoms by electrical discharges (lightning). At altitudes of 25 to 40 km, however, its concentration increases substantially. /6
Oxygen is dissociated by the ultraviolet rays of the sun into its component atoms. The latter, by recombining with the remaining diatomic molecules of oxygen, form triatomic oxygen, called ozone. Its maximum concentration occurs between 20 and 25 km and drops sharply from these altitudes, both upward and toward the earth's surface. The total concentration of ozone is extremely small by comparison with other gases. Reduced to standard conditions (760 mm Hg at 0°C), it would be equivalent to a layer with a total thickness of only 0.3 cm. Nevertheless, the importance of ozone is enormous. Should suddenly it vanish from the atmosphere, all life on earth (including plants) would be incinerated by ultraviolet rays. Also extremely important is the fact that

ozone, by absorbing much of the sun's ultraviolet radiation, as well as thermal radiation (long-wave radiation) escaping from the earth, heats the stratosphere at altitudes of 30-50 km. This heating is particularly substantial over the polar regions during the summer, when the greatest amount of ozone is formed over these regions. This has a considerable effect on the distribution of pressure, temperature and wind at the stated altitudes.

It has been established that the concentration of ozone is maximum in the spring and minimum in the autumn, and the annual amplitude increases with latitude. Aperiodic variations in ozone concentration have also been detected in connection with variations in the synoptic position in the troposphere. Tropical air masses contain less ozone than do arctic air masses; ozone concentration is generally smaller in the leading parts of depression than in the trailing; in particular, the passage of a warm front is accompanied by a reduction of ozone concentration, and the passage of a cold front, by an increase. There is a relationship between ozone concentration, solar activity and state of the earth's magnetic field. The ozone concentration decreases when solar activity increases.

The layer of the stratosphere in which occur the photochemical processes of ozone formation and where ozone concentration is particularly high, is called the ozonosphere.

1.2. Standard Atmosphere

The various meteorological elements have a tremendous influence on aerodynamic and performance characteristics of aircraft, on the readings of navigation instruments, engine operation, etc. This is particularly true of modern jet-propelled high speed aircraft. Meanwhile, the actual distribution of meteorological elements in the atmosphere is characterized by a rather considerable variability. In this connection, in the design of aircraft and /7
instruments and also for the purpose of obtaining comparable results from their testing under various meteorological conditions, it is customary to reduce them to certain standard conditions. An atmosphere with such a distribution of basic meteorological elements is called a standard atmosphere (SA). The parameters

and physical characteristics of SA are found in consideration of long-term (over many years) mean values of meteorological elements.

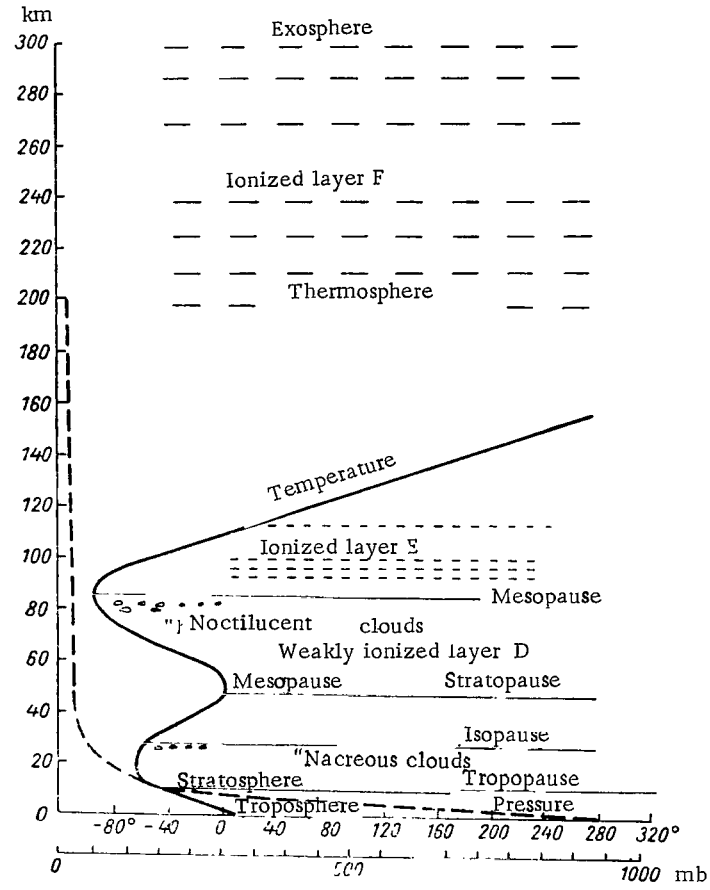


Figure 1.1. Structure of Atmosphere up to 300 km.

Naturally, as new data on the physical properties of the atmosphere are accumulated and as aircraft are developed and improved, the parameters of the standard atmosphere should also be corrected. Until 1964 the USSR employed SA designed for altitudes up to 30 km. In 1964 a new standard atmosphere was introduced (GOST 4401-64). /8

As parameters and physical constants, SA-64 incorporated temperature, pressure, density of the air, relative (in relation to their values at sea

level) air pressure and density, molecular weight[1], speed of sound, viscosity (dynamic and kinematic), free fall acceleration and molecular mean free path.

The temperature and molecular weight of the air were used as the initial parameters for calculation of the tabular values. With respect to the character of temperature change the atmosphere is divided into 11 layers within which molecular temperature T_M changes linearly with geopotential altitude.

The molecular weight of air in the SA is constant up to 95 km, but decreases above that altitude.

The physical characteristics of SA-64 at sea level are:

atmospheric pressure p = 1,013.25 mb = 760 mm Hg,
temperature T_0 = 15°C = 288.15°K,
molecular weight of air M_0 = 28.966.

1.3. Layers of the Atmosphere

The atmosphere is divided vertically into several layers, called spheres. Each of these layers is distinguished by a unique temperature change in the vertical, and also by characteristic features of other physical properties: composition, structure, electrical characteristics. The mean altitudes of the basic and transition layers of the atmosphere are indicated in Table 1.1. The vertical cross-section of the atmosphere is shown schematically in Figure 1.1, /9 and the curves of change of temperature and pressure with height are also shown there in accordance with the standard atmosphere.

We will examine the features of the basic and transitional layers, bearing in mind that the chief indicator of each of them is the corresponding temperature characteristic.

Troposphere

The troposphere is the bottom most part of the atmosphere, extending from the earth's surface to the altitude of 8-10 km in the northern latitudes, to

[1]Molecular weight is defined as the weight of a molecule expressed in units equal to 1/16 the weight of the hydrogen atom; equal to the sum of atomic weights of the atoms of which the molecule is composed.

a height of 10-12 km in the middle latitudes, and up to 16-18 km over the equatorial regions. In this comparatively thin layer, as already mentioned, is concentrated about 90% of the entire mass of the atmosphere. Therefore the air here is densest.

Table 1.1. Principal Transition Layers (Spheres) of the Atmosphere

Layer (sphere)	Average altitude of upper and lower boundaries (km).	Transition layer
Troposphere	0 - 11	
Stratosphere. . . .	11 - 50	Tropopause
Mesosphere	50 - 90	Stratopause
Thermosphere. . . .	90 - 800	Mesopause
Exosphere	above 800	Thermopause

Solar radiation has extremely little direct influence on tropospheric air. It consists basically in heating of the earth's surface, from which heat is subsequently transmitted to the atmosphere in the form of long-wave radiation. Most of this heat, naturally, will be absorbed by the lower layers of the troposphere, and a small part by the top layers of the troposphere. As we know, the characteristic feature of the troposphere is a reduction of temperature with altitude, which averages 6.5° per km. Here, on individual days, particularly during the winter, in the lowermost layers of the atmosphere the vertical temperature gradients can vary within broad limits from positive values to negative (temperature inversions).

Nearly all the water vapor, condensation of which produces clouds and precipitation, exists in the troposphere, in contrast to the upper layers. Air masses and fronts, along with cyclones and anticyclones, form in the troposphere. The troposphere is also the dustiest part of the atmosphere. It is more accessible to investigation and therefore has been investigated more thoroughly than the other layers of the atmosphere. It is precisely here that the basic processes, which reveal themselves as showers, thunderstorms, hurricanes, snow showers and other phenomena, occur.

Thorough analysis of the troposphere reveals in it the following three layers:

1. Boundary layer--layer of the atmosphere between the earth's surface and an altitude of 1,000-1,500 m. Under the influence of the earth's surface and turbulent exchange, well-defined daily changes of meteorological elements occur within it, the wind velocity in this layer increases, as a rule, with altitude, the wind velocity vector near the earth's surface declines at an average of 30-35° from the isobars toward low pressure, and the angle of declination, due to reduced friction, decreases practically to zero with altitude.

The lowest part of the boundary layer, up to 50-100 m, is known as the surface layer; the direct thermal influence of the earth's surface (heating or cooling) is strongest in the surface layer. /10

2. The middle troposphere, between 1-1.5, up to 6-7 km, is the layer of the most active cloud and precipitation activity.

3. The upper troposphere, from 6-7 km to the tropopause, is distinguished by the strongest winds--the jet streams.

The layer of the troposphere above 1,500 m is often called the free atmosphere. Discounting the mountainous regions, roughness (friction) of the earth's surface has practically no direct influence here.

Tropopause

The tropopause is the transitional area between the troposphere and the stratosphere. The lower boundary of this layer over the polar region averages 8 km, in the middle latitudes--about 10 km, and in the tropics--about 18 km. The tropopause is lower during the winter than during the summer. The mean temperature at the level of the tropopause over the poles is about -50°, and from -75 to -80° over the equator. The level of stable transition to the isotherms or to small (in terms of absolute value) vertical temperature gradients ($\gamma \geqslant 0.2°/100$ m) is used as the upper boundary of the troposphere for processing and coding of radiosonde data of the atmosphere.

The thickness of the tropopause can vary considerably. Sometimes there is a pronounced transition from troposphere to stratosphere, and then the tropopause represents a sort of discontinuity boundary between the air masses of the troposphere and stratosphere. More often the tropopause is a transitional layer measuring from a few hundred meters to several km in thickness with a complex thermal structure. Here, the following fact is noteworthy. At high latitudes in the winter, due to considerable radiative cooling of the upper layers of the troposphere, the tropopause seems to recede upward and is weakly detected only at altitudes of 18 and 20 km, and sometimes higher. Thus the tropopause more or less vanishes. The so-called tropopause discontinuity occurs in the southern latitudes (30-40°). North of these latitudes is located the polar tropopause at an average altitude of about 10 km, and to the south the tropic tropopause at altitudes of 14-16 km.

The location of the tropopause changes continually, both in time and in space, and this is closely related to the movement of warm and cold air masses in the troposphere, movement and evolution of cyclones and anti-cyclones. The basic principles of these relations consist in the following:

a) The regions of lowest altitudes of the tropopause coincide with cold centers on the chart OT $\frac{500}{1000}$, and the regions of maximum altitude with warm centers;

b) During the period between the development and filling stages of a cyclone the tropopause descends continuously over the center of the cyclone; during the cyclone filling period the tropopause gradually rises over it;

c) From the moment of formation of an anticyclone to the moment of maximum development the tropopause continuously rises over the center of the anticyclone; during collapse of the anticyclone the tropopause slowly descends over it; /11

d) As a warm front approaches a given point the tropopause above that point rises; accordingly, as a cold front approaches the tropopause descends.

By examining such periodic fluctuations of tropopause altitude related to the passing and evolution of cyclones and anticyclones it can be concluded

that as the tropopause rises its temperature decreases and, conversely, as the tropopause descends its temperature rises.

Stratosphere

The stratosphere is the layer of the atmosphere lying above the troposphere and extending to an altitude to the order of 50 km.

As seen in Figure 1.1, the temperature in the lower part of the stratosphere, i.e., from the tropopause to an altitude of 20-25 km, varies little with altitude, but then it increases to an altitude of about 50 km, even acquiring positive values at this level. Thus the stratosphere can be divided into two layers: in the first (bottom) layer the temperature change with altitude is nearly isothermic, but in the second (top) layer temperature change is of the inversion variety. The transition layer between them, according to V. R. Dubentsov's theory [5] is known as the isopause. The above-mentioned temperature change with altitude is evidence of the fact that vertical movement of air in the stratosphere is either nonexistent or very weak compared to that in the troposphere.

The chief cause of substantial temperature rise in the upper half of the stratosphere is, as mentioned above, the higher ozone concentration. Cyclones and anticyclones also occur in the stratosphere, but they have an entirely different origin and, compared to tropospheric processes, are more stable in time and in space.

As regards the problem of clouds in the stratosphere, since the amount of water vapor, like the amount of condensation nuclei, in the stratosphere is exceedingly small, the formation of clouds in these layers is improbable. In rare cases cumulonimbus clouds in the southern latitudes penetrate from the troposphere into the lower part of the stratosphere. Along with this, at extremely low temperatures vapor can condense in certain cases even in the stratosphere, and then streaks of ice crystals form, even during slight vertical movements, which can spread over great distances. The thickness of these formations is small, and so is their density. Therefore they are not visible from the surface of the earth. However, aviators, as a rule, notice these formations and see them as a cloud in the stratosphere. Sometimes the thick- /12

ness of such formations is great enough so that they can be seen from the earth with the appropriate illumination. These are the same as nacreous clouds and occur almost exclusively in the winter (from November to March) at altitudes between 20 and 30 km. The maximum occurrence is in January. As a rule they form at an altitude of about 25 km, i.e., near the isopause.

Nacreous clouds are usually seen in Scandinavia, Alaska and in the Caucasus, i.e., in mountainous regions. The formation of such clouds is usually related to orographic wave motions in air flows of favorable thermal structure. Their appearance can also be related to atmospheric turbulence in the delta of stratospheric frontal zones that form in the winter above the northwestern regions of North America and Europe.

The transition layer between the stratosphere and the mesosphere, located at an altitude of about 50 km, is called the stratopause.

Mesosphere

The mesosphere is the layer of the atmosphere located from 50-90 km. Its distinguishing feature is reduction of temperature with altitude from about 0° at an altitude of about 50 km to -80 to -90° at about 90 km. Such a temperature change indicates that the mesosphere should be distinguished by high turbulence. Also evidence for this, in particular, is the fact that luminous clouds, called silvery (cirrus) clouds are sometimes seen at altitudes between 80 and 90 km, i.e., near the upper boundary of this layer. They generally appear in the northern sky on summer nights and move generally from east to west at 40-80 km per hour.

Thermosphere

This layer of the atmosphere is located above 90 km and is characterized by a continuous temperature rise. The temperature presumably reaches 150 to 300° at an altitude of about 160 km.

The transition layer between the mesosphere and the thermosphere is called the mesopause.

The layer of the atmosphere above 80-90 km contains considerable quantities of electrically charged fine particles, molecules or atoms of atmospheric gases, called ions. Therefore the above-mentioned layers are also called the ionosphere, consequently the thermosphere essentially coincides with the ionosphere. The latter term, however, is used only in those cases when discussion concerns the electrical properties of the atmosphere. These layers of the atmosphere are marked by high electrical conductance, which at the altitude of 100 km, for instance, is several billion times greater compared to the electrical conductance of air on the earth's surface.

The ionization of the upper layers of the atmosphere is the result of the action of the ultraviolet part of solar radiation on the atmosphere, and also of small charged solid particles, called corpuscles, discharged by the sun.

Ionization has a tremendous effect on the propagation of radio waves, which are refracted, reflected or absorbed in the ionosphere. These phenomena have been thoroughly investigated, and this makes it possible to employ radio waves for direct analysis of the ionosphere. It has been found that individual /13 layers of the atmosphere at great altitudes have a different degree of ionization. Two such layers are most discernable: the E layer, located at altitudes between 80 and 100 km, with maximum ionization at about 100 km, and the F layer, at about 200 km. In addition there is yet another layer of weak ionization, the D layer, located in the mesosphere at 50-65 km.

The state of the ionosphere is not constant, but changes continuously during the year and day, and also as a function of changes in solar activity. In this connection the conditions of radio reception also change. Thus, for instance, in the daytime the degree of ionization of the lower layers of the ionosphere increases under the influence of solar radiation, and radio transmitters are more stable on shortwaves (10-50 m). At night, in connection with attenuation of ionization, longer radio waves are used. The disruption of shortwave radio communications is particularly strong during periods of so-called magnetic storms, which usually occur in the high latitudes.

Exosphere

The exosphere is the outermost layer of the atmosphere, located above 800 to 1,000 km. Here, under conditions of an extremely rarefied atmosphere, gas molecules travel at extremely high velocities (in excess of 11 m/sec). Consequently, these molecules, in overcoming the force of gravity, can escape into interplanetary space.

How far out does the earth's atmosphere extend? This question can be answered first of all from the standpoint of theoretical considerations based on the hypothesis that at the upper boundary of the atmosphere the centrifugal force of the earth's rotation should be equalized by the earth's gravity. Calculations indicate that this theoretical boundary of the earth's atmosphere is 28,000 km above the poles and 42,000 km above the equator. These data, however, are only of theoretical interest. Presumably, at these altitudes only isolated gas molecules, separated by the space of 1,000 km, will be encountered.

From the practical point of view it is essential to know the altitude of those layers of the atmosphere whose density is measurable and in which occur physical phenomena that can be observed. Among such phenomena that occur at great altitudes are Auroras Borealis. This optical phenomenon is closely related to ionospheric magnetic storms and represents the glowing of the gases of a rarefied atmosphere under the influence of bombardment by electrically charged particles discharged by the sun. It has been established that the upper boundary of an Aurora Borealis is 1,000 to 2,000 km, which until recently was also regarded as the upper boundary of the atmosphere.

The launch of the first Soviet artificial earth satellite on October 4, 1957 and subsequent launches, both by the USSR and the USA, together with the launching of space missions presented scientists with vast possibilities for additional thorough and comprehensive analysis of the atmosphere. The scientific results obtained have already shed new light on many questions concerning the physics of the upper atmosphere. We will mention some of these results. /14

The mean density of the atmosphere at the perigee altitudes of Soviet satellites (225-228 km), calculated on the basis of their deceleration, was

$3 \cdot 10^{-7}$ g/m^3. This is about 5 to 10 times greater than the values obtained earlier on the basis of the processing of the data of rocket observations. It was discovered that the density of the atmosphere at a given altitude is not constant. It changes particularly sharply on days when magnetic storms occur on the earth as a result of sun flares. It was also found that the density depends on the time of day. The density of the atmosphere is much greater over the illuminated side of the earth than at the same altitude over the dark side. Atmospheric density turned out to be less in the equatorial region than in the middle latitudes at the corresponding altitudes.

On the basis of data on deceleration of earth satellites it is presently considered that the upper boundary of the atmosphere--the geocorona--extends out to 20,000 km. It is presumed that the density of the atmosphere at these altitudes is equal to the density of interplanetary space.

Investigation of the ionosphere by "illuminating" it from above with radiowaves emitted by transmitters aboard satellites revealed that regions with a high ion concentration are located considerably higher than previously thought.

Artificial earth satellites made it possible to discover the earth's radiation belts. They represent zones of high concentrations of charged particles caught in a trap created by the earth's magnetic field. These are zones of high intensity radiation.

It has been established that the outer radiation zone extends in the equatorial plane from 14,000 to 50,000--55,000 km from the earth's surface. In the range of latitudes of 50-70° the outer zone approaches 270-300 km. The boundary and peak radiation of this zone vary in time.

The inner radiation zone is located 600 to 5,000 km from the earth's surface. In contrast to the outer zone, the inner zone is more stable in time and its radiation harder, and consequently more dangerous to the crews of long-term space missions.

One of the important purposes of investigations at the present time is the solution of the problem of the influence of the radiation belts on the temperature of the upper layers of the atmosphere located below 1,000 km.

/15

CHAPTER 2

SEASONAL FEATURES OF CIRCULATION IN THE STRATOSPHERE

2.1. General Information

Systematic studies of the features of circulation in the stratosphere were undertaken in many countries of the world in 1957 in connection with the completion of the International Geophysical Year program.

The expansion of the aerological network of stations, elevation of the atmospheric sounding ceiling, launching of meteorological rockets and artificial earth satellites, greatly assist in understanding the essence of circulation in the stratosphere. Studies conducted over the last ten years have shown, in particular, that contrary to previous notions, pressure, temperature and wind, like atmospheric circulation in the stratosphere as a whole, experience considerable changes, although to a lesser extent than in the troposphere. These changes occur both as functions of seasonal changes and within each season, particularly during the cold half of the year.

Seasonal circulation features are governed principally by the thermal regime of the stratosphere, which depends, in turn, on the solar illumination of the hemisphere during the various seasons of the year and on the absorptive capacity of ozone, and also partially on the character of thermal effects of the underlying surface (the oceans and continents).

As we know, the sun rises once a year at the North Pole--the 21st of March (vernal equinox) and, without setting for six months, describes daily a complete circle, which simultaneously rises to an ever higher altitude until the 22nd of June (summer solstice). Then the sun drops to the horizon and sets on the 23d of September (autumnal equinox) for six months. Thus the polar region is illuminated by the sun for the six warm months, i.e., from the 21st of March to the 23d of September (186 days), and during this period /16
the stratosphere over the given region is heated.

During the cold half-year (179 days) darkness prevails on the North Pole and during this time the stratosphere cools considerably. Since a high geopotential corresponds to a warm air mass and a low geopotential to a cold air mass, anti-cyclonic circulation prevails in the stratosphere during the warm period, and the cyclonic during the cold period.

Maps of baric topography of the 200, 100, 50, 30, 20 and 10 mb levels are generally used in the investigation of processes in the stratosphere. In the given case the 200 mb level has a comparatively small value, since for most regions of the earth this level is located in the troposphere. Consequently, the geopotential and temperature fields at the 200 mb level reflect primarily tropospheric processes. The temperature and geopotential fields on the 100 mb level (15-16 km) are notably rearranged, the horizontal gradients of these elements are reduced and the winds subside.

The most notable reorganization of the geopotential and temperature fields occurs at the 50 mb level and above. The results of analysis of maps for the individual levels of 50, 30 and 20 mb are very close to each other in the sense that they explain general characteristics of the development of stratospheric processes. They are all located in the middle part of the stratosphere and all satisfactorily reflect processes that occur in the layer between the tropopause and isopause; here the 30 mb level often falls in the isopause layer itself. Temperature differences at the 100 and 50 mb levels, compared to the 30 mb level, usually fall in the range of $0 \pm 4°$. It was decided in this work to limit the discussion to a description of the evolution of temperature, geopotential and wind fields in the stratosphere on the basis of the data for a single level, specifically according to AT_{30} maps which corresponds to elevations of 20-24 km. Other levels will also be examined in the description of individual cases. In the analysis of the high level maps, however, we will encounter considerable difficulties, the reasons for which, as pointed out by V. R. Dubentsov [5], are the following.

First, the number of observations drops off sharply with altitude; the higher the isobaric level, the smaller the number of probes that reached the corresponding level.

Second, most countries employ their own radiosonde systems for aerological observations. Most types of radiosondes give comparable data up to the 200 mb isobaric level, but beyond that level there are considerable differences in instrument readings. The largest errors occur due to different heating of the instruments during the daytime. These are called radiation errors. They depend on instrument design, height of the sun, density of the air, vertical velocity, cloud cover and reflectivity of the instrument surface. According to the data of international comparison of radiosondes, conducted in 1956 in Payern [Switzerland], the Soviet radiosonde RZ-049 had the following radiation error: about 2° at an altitude of 10 km, about 3° at 17 km, and 7-8° at 23 km. /17

New and improved types of radiosondes--A-60, RKZ-2, have appeared in recent years, in which connection, as indicated by experience in the analysis of high level maps at the USSR Gidromettsentr [Hydrometeoerological Center], the temperature data on these maps are, from a practical standpoint, quite comparable and satisfy practical requirements. The exceptions are cases when no high-altitude air temperature data are available for some territory. This happens when radiosondes, during their ascent, enter a region of very low air temperature. Because of insufficient resistance of radiosonde cases to cold, they break apart at some height and we are thereby deprived of the necessary data. This usually occurs at altitudes where the air temperature is below -70°. Data concerning wind direction and velocity are not too bad, but the situation with geopotential values is much worse. Arriving at a more acceptable picture of the baric field at high altitudes requires very careful analysis of the wind field based on the theoretical principles of the relation between wind and the pressure (geopotential) field.

Before proceeding to the problem of the basic types and features of stratospheric circulation it is useful first to have a general representation of the temperature regime in the stratosphere during the different seasons of the year and in the various geographic regions. For this purpose we will turn to Table 2.1 and 2.2, in which are listed the mean air temperatures at the 30 and 50 mb levels, respectively, based on data for 1966 and 1967. Several

Eurasian points, located latitudinally from west to east (Table 2.1) and meridionally from north to south (Table 2.2) are used. These data are given only for January, April, July and October. Although data of specific observations may differ substantially from the mean air temperatures presented here, taken as a whole they nevertheless correctly reflect the fundamental principles of the temperature regime in the middle stratosphere.

TABLE 2.1. MEAN AIR TEMPERATURE (°C) AT THE 30 AND 50 mb LEVEL AT POINTS LOCATED FROM WEST TO EAST.

Point	Jan.	Apr.	July	Oct.
30 mb				
London	—62	—56	—47	—57
Moscow	—63	—56	—45	—57
Omsk	—62	—56	—46	—55
Irkutskh	—55	—51	—46	—52
Khabarovsk	—45	—48	—48	—50
50 mb				
London	—61	—56	—51	—58
Moscow	—62	—57	—49	—58
Omsk	—61	—56	—48	—57
Irkutskh	—56	—54	—49	—54
Khabarovsk	—46	—53	—52	—51

TABLE 2.2. MEAN AIR TEMPERATURE AT 30 AND 50 DB LEVELS AT POINTS LOCATED FROM NORTH TO SOUTH.

Point	Jan.	Apr.	July	Oct.
30 mb				
North Pole	—77	—47	—38	—64
Arkhangel'sk	—70	—54	—42	—59
Moscow	—63	—56	—45	—57
Rostov-on-Don	—62	—56	—48	—55
Tblisi	—55	—55	—50	—55
50 mb				
North Pole	—74	—49	—38	—61
Arkhangel'sk	—67	—56	—44	—58
Moscow	—62	—57	—49	—58
Rostov-on-Don	—58	—56	—52	—58
Tblisi	—58	—58	—56	—58

Analysis of each of the given tables leads to the following conclusions.

As seen in Table 2.1, the lowest temperatures at the 50 and 30 mb levels occurred in January, i.e., during the cold season; here the temperature is somewhat higher in the west to east direction, particularly in the Irkutsk--Khabarovsk segment. The highest temperatures occur in July; here the temperature at the 30 and 50 mb levels is much lower over extreme Western Europe (London) and Far East (Khabarovsk) than over the continental stations at the same latitudes.

During the transition seasons (April and October) the temperature distri- /18
bution in the latitudinal direction does not experience notable variations and remains approximately the same, about -55°.

The temperature at the 50 mb level is somewhat lower in most cases than at the 30 mb level (the mean deviations fall within the limits 1-5°); the greatest negative deviations occur during the warm period of the year (July) /19
and the smallest in the spring (April); in individual cases the temperature difference reaches 7-8°.

It follows from Table 2.2 that a general temperature rise occurs from the North Pole to the South, approximately to 40°N latitude in the autumn (October) and winter (January); in the spring (April) and summer (July) a general temperature reduction occurs from the North Pole to the south.

The greatest temperature increase occurs over the Pole between January and April, amounting to about 30° at the 30 mb level and 25° at the 50 mb level; moving southward, the temperature rise diminishes, and at about the 40° latitude the temperature remains practically constant.

3. The temperature reduction over the Pole between July and October is also maximum, although it is less in absolute value than the temperature rise between winter and spring.

We will proceed to the problem of the basic types of circulation in the stratosphere.

2.2. Summer Stratospheric Anticyclone

We will examine how atmospheric processes develop and manifest themselves in the stratosphere during the warm period of the year, starting with April.

The presentation of this problem is based chiefly on research results [5 and 14].

The average AT_{30} chart illustrated in Figure 2.1 pertains to April, i.e., to the transitional season. It shows that the baric and thermic fields in the stratosphere are characterized during this period by comparatively small horizontal pressure and temperature gradients. The polar region and the zone of middle latitudes are still under the influence of winter cyclonic, but by now substantially weakened, circulation. Farther south, at the 40-20°N latitudes, there is anticyclonic circulation, determined by several individual anticyclones. Thus, westerly winds, with velocities of the order of 10 m/sec at level AT_{30}, still prevail in the spring over most of the Northern Hemisphere. The zone of transition from westerly circulation to easterly is located at the 20-30° latitudes. Here, as in Figures 2.2, 2.4, 2.6 and 2.7, it is denoted by the small circles.

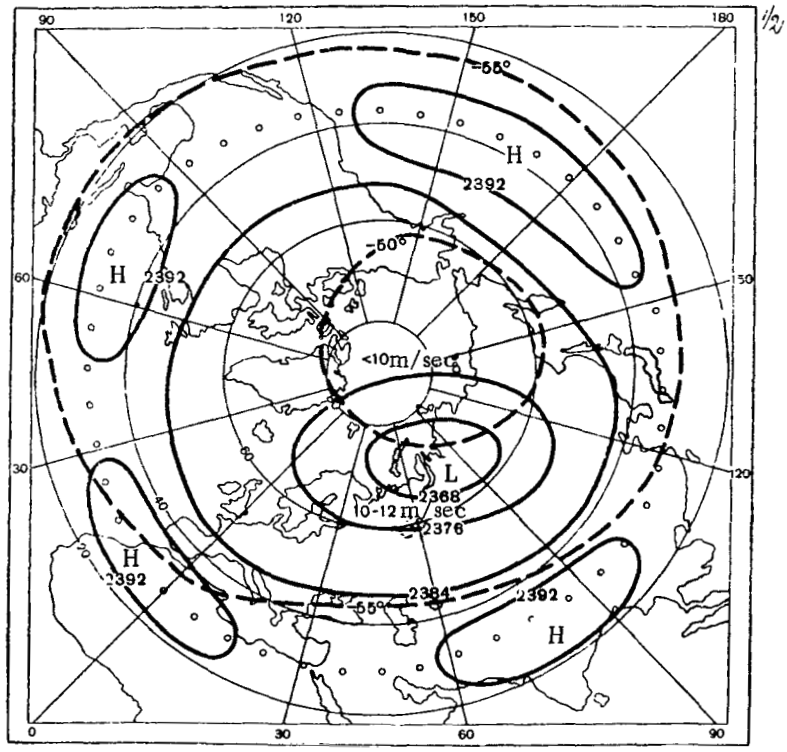


Figure 2.1. Chart of Mean Geopotential, Temperature and Wind at 30 mb Level. April.

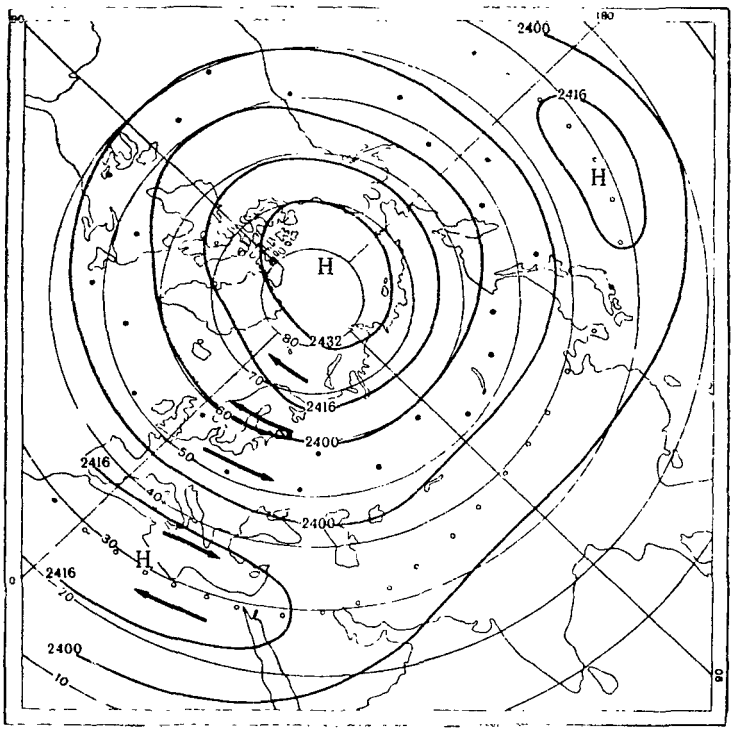


Figure 2.2. Average AT_{30} Chart. Third Decade of May 1958.

It should be pointed out, however, that the horizontal temperature gradient at the given level now extends from the North Pole to the Equator: the mean temperature for the month over the polar region is -50°, and in the southern latitudes (between 40 and 20°) -55°. With such a temperature distribution thermal wind can result in substantial variations in circulation in the month of April.

The formation of the anti-cyclone as an independent baric event usually /20
occurs over the polar region between the middle of April and middle of May.

Analysis of the AT_{30} chart for the third decade of May 1958 indicates (Figure 2.2) that the entire polar region, and also the northern regions of the hemisphere, approximately to 60°N latitude, are under the influence of anti-cyclonic circulation during this period. The center of the polar anti-cyclone migrates during this period in the region of the Canadian Archipelago.

The average decade altitude of the 30 mb level at the center of the anticyclone is about 2,432 decameters.

In connection with the expansion of the polar anticyclone the zone of cyclonic circulation is forced southward, and after a certain period of time it stabilizes between the 60-40°N latitudes with an average geopotential less than 2,400 decameters. In this comparatively narrow zone the wind changes direction from east to west, as shown schematically in Figure 2.2 by the arrows. Still farther south, approximately between the 40 and 25°N latitudes, there is a second anticyclonic region with a maximum geopotential of about 2,420 decameters. South of the 25th parallel is the transition of low pressure into the equatorial zone, where the altitude of the 30 mb level is about 2,380 decameters. As already mentioned, the chain-like line that passes through the tropical zone is the boundary between westerly circulation north of this line and easterly circulation south of it. /21

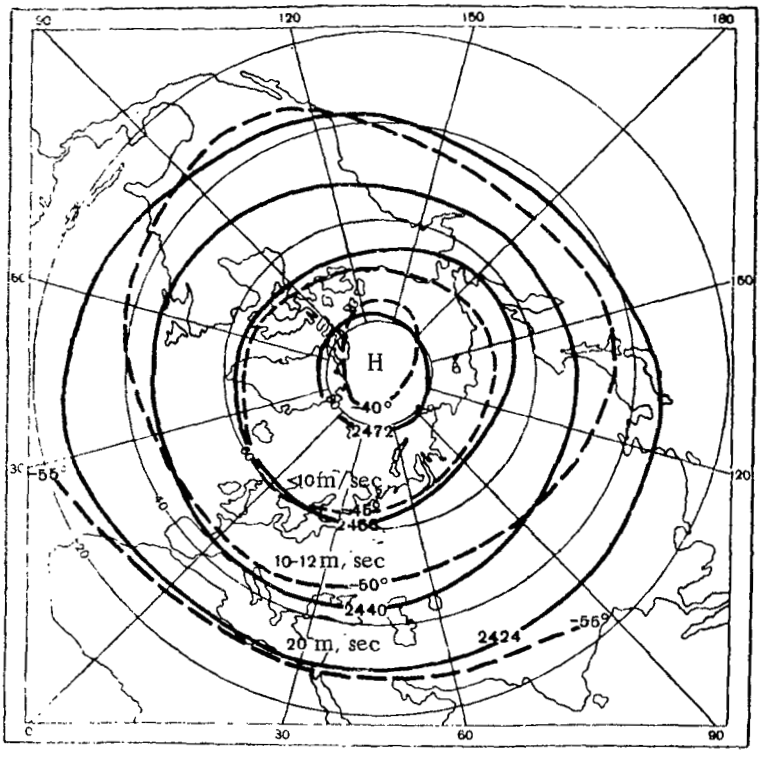


Figure 2.3. Chart of Mean Geopotential, Temperature and Wind at 30 mb Level. July.

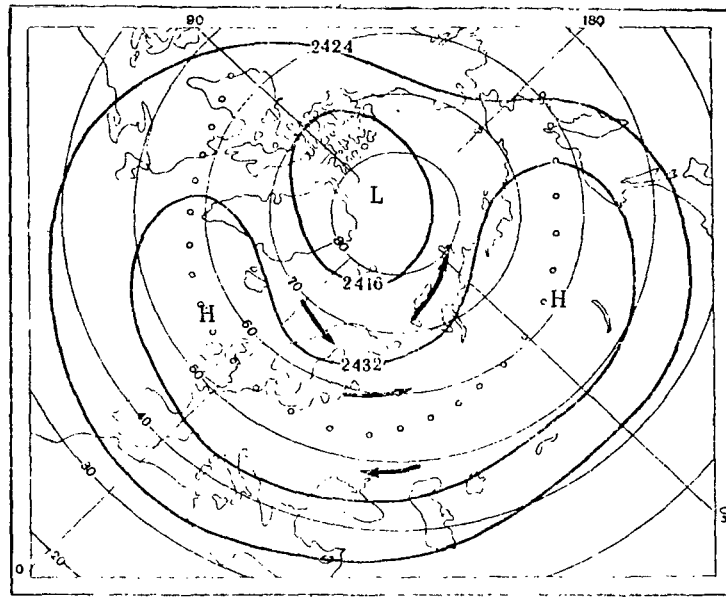


Figure 2.4. Average AT_{30} Chart. Third Decade of August 1958.

As the polar anticyclone intensifies, its central part shifts from the Canadian archipeligo (where it was formed) to the North Pole. Simultaneously the influence of this anticyclone spreads to the southern and middle latitudes, while the cyclonic circulation that previously existed here rather rapidly fades away. By the middle of June the polar anticyclone combines, in a manner of speaking, with the subtropical high pressure zone and in the presence of an equatorial depression the horizontal geopotential gradient over the entire hemisphere extends from north to south. In other words, in the second half of /22
June a homogeneous anticyclonic field of circulation is established over the entire hemisphere in the form of a circumpolar eddy, in which the atmosphere moves generally from east to west. On certain days, however, the baric field at these altitudes takes on a character of shifting shallow troughs and ridges, and therefore the winds are southeasterly in some regions and northeasterly in others.

During most of July the stratospheric anticyclone still continues to intensify, reaching its maximum development (≈2,480 decameters) during the

third decade of this month (Figure 2.3). The region of maximum temperature (about -40°) coincides here with the central part of the anticyclone; at about 30° latitude the air temperature at the 30 mb level is about -55°. In the /23
northern latitudes the velocity of the easterly winds is less than 10 m/sec, 10-12 m/sec in the middle latitudes and of the order of 20 m/sec in the tropics. The strongest easterly winds are concentrated between the 30 and 15° N latitudes.

Starting on the 22nd of July the anticyclone gradually begins to weaken. Nevertheless, for yet another month, i.e., up until the third decade of August, the anticyclone continues to exist as an independent baric formation with its center near the pole, and only in the third decade of August does substantial reorganization of the AT_{30} field take place. This reorganization consists in the appearance of cyclonic formations above the polar region in a path across the Canadian Archipeligo (Figure 2.4) and a reduction of temperature. The anticyclonic region, indeed, occupies only the middle latitudes. The axis of this anticyclonic system passes over North America and the Atlantic Ocean along about the 50th parallel, over Europe approximately at 55° latitude, and then over Asia along about 60° latitude.

Such are the principal features of evolution of the summer stratospheric anticyclone.

Changes in altitudes of the 30 mb level over the polar region from April through August, 1958 and from April through July, 1959, are illustrated in /24
Figure 2.5. Both curves show that between April and the summer solstice (22nd of July) the isobaric surface rises rather rapidly. This can be explained by progressive heating of stratospheric air as a result of ozone absorption of ultraviolet radiation. During the following month the isobaric surface stays at about the same level, but then falls rapidly. It is noteworthy that the beginning of the formation of the polar anticyclone in the spring (April--May) like its disappearance toward the end of August, corresponds to the time when the geopotential at the 30 mb level in the region of the pole reaches 2,400 to 2,420 decameters. Changes in the altitudes of the 25 mb level over the polar region during the period July-August, 1957 and April-August 1958, are illustrated in the same figure.

These curves are quite similar to those corresponding to the 30 mb surface (Figure 2.4), which indicates that the process of anticyclogenesis is quite the same at the 30 and 25 mb levels.

The above-described principles of evolution of the baric field in the stratosphere are typical of the warm time of the year, although there may be considerable deviations from these principles during individual years.

In analyzing daily AT_{30} maps, for instance, we often encounter a situation /25 where certain indications of cyclonic circulation over limited regions are detected on the general background of an intensifying anticyclonic circulation. This is manifested either in changes in wind direction or in changes in altitudes of an isobaric level. These facts cannot be ignored, since they usually occurred during certain processes in the middle latitudes, in particular, during an invasion of a cold air mass in the troposphere, caused by a deepening of the tropospheric trough in the given region, which can result in certain violations of circulation in the stratosphere as well.

Comparison of the processes of one year with others also indicates that deviations from the above-described model are notable during individual years. In particular, analysis of the curves of change of the altitudes of the 30 mb level over the polar region (Figure 2.5) reveals a substantial difference in the absolute values of altitudes during the period from the 10th of April through the 20th of May of the two stated years. In 1959 the elevations of the 30 mb level were much greater than during the same period of 1958, particularly in April. In the spring of 1959, obviously, additional factors came into play, which facilitated more active heating of the air in the stratosphere, and in this connection, elevation of the 30 mb level. This was manifested in the formation of a polar stratospheric anticyclone between the 10th and 15th of April, 1959, i.e., almost a month earlier than in 1958 (10th of May). Having formed, as in 1958, over the American sector of the arctic, this anticyclone spread over Europe much earlier than in the previous year.

We might point out here that in eastern Siberia, April of 1959, in contrast to April of 1958, was characterized by a long-term existence of a vast stable depression at the above-mentioned level.

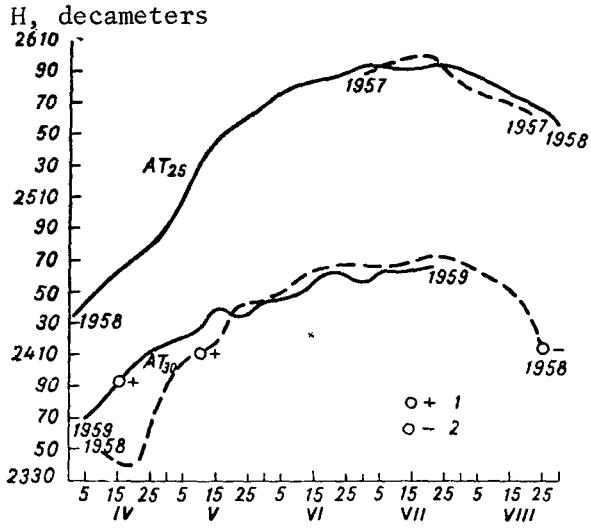


Figure 2.5. Change of Altitudes of 30 and 25 mb Levels over the Region of North Pole. 1, Origin of anti-cyclone; 2, Destruction of anti-cyclone.

This all is evidence that the structure of the baric field during the same seasons may differ from year to year, even at high altitudes.

Anticyclonic circulation in the stratosphere does not begin right at the transition from the troposphere to the stratosphere, but at some elevation above the tropopause.

There is a transitional layer that separates the stratospheric easterly anticyclonic circulation from the underlying westerly cyclonic circulation. This layer is characterized by weak, unstable winds. Its thickness varies tremendously, from 1 to 5 km, but its lower boundary is placed at the level at which the stable easterly regime begins in the stratosphere. This layer is discussed in greater detail in Chapter 3, and also in I. V. Bugayeva's work [1].

2.3. Winter Stratospheric Cyclone /26

Starting with the last days of July the polar stratospheric anticyclone begins gradually to weaken and to yield to cyclonic circulation. However, during August and the first half of September, i.e., almost until the autumnal equinox, cyclonic circulation over the polar region is weak. The horizontal temperature and the geopotential gradients during this time are negligible. Consequently, very weak winds, unstable in direction, prevail in the stratosphere. Individual cyclonic eddies that appear are also very unstable, both in time and in space.

After the autumnal equinox radiative cooling of the atmosphere intensifies, resulting in further reduction of temperature and geopotential. As seen from examination of the average geopotential and the temperature map for October

(Figure 2.6), this month is characterized by distinct cyclonic circulation, the center of which is near the pole, and which encompasses much of the northern hemisphere. The horizontal temperature and geopotential gradients extend from /27
south to north over most of the northern hemisphere. The west winds, although somewhat strengthened in the polar basin as compared to the middle latitudes, but do not remain strong (on the order of 10 m/sec). The east winds south of the 30th parallel feature maximum velocity. The region of relatively warm stratosphere over the Pacific Ocean, near the Asian coast, is well defined. At the same time, the temperature above the polar region falls about 26° below the July temperature and averages about -64°.

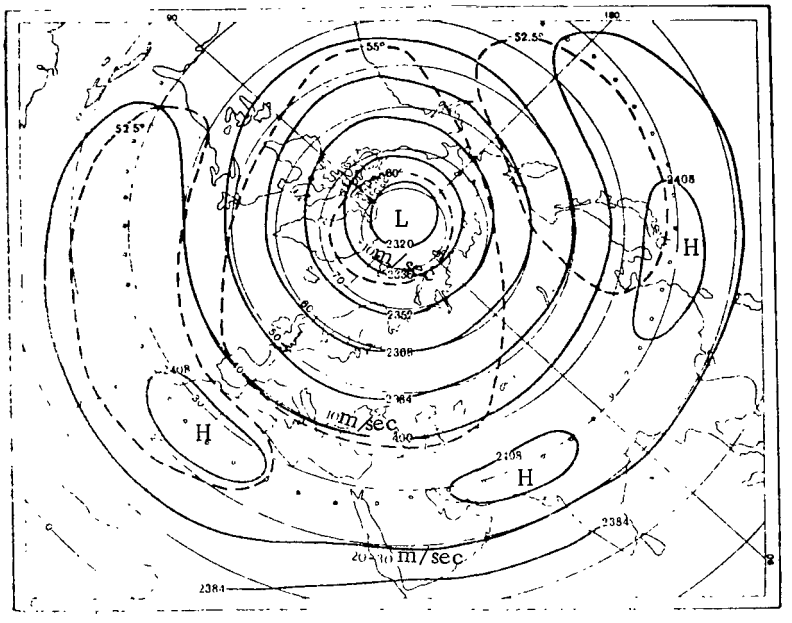


Figure 2.6. Chart of Mean Geopotential, Temperature and Wind at 30 mb Level. October.

By the hibernal solstice (22 December) radiative cooling of the atmosphere increases, and therefore cyclonic circulation in the stratosphere reaches its maximum development by this time.

The AT_{30} map that reflects the average baric field at altitudes of 21 to 25 km for January, i.e., for the very middle month of the cold season, is illustrated in Figure 2.7. We see that a stratospheric cyclone appears on the map as an extensive system of isohypses, which outline two troughs. One extends into eastern Asia, and the other into eastern North America. The minimum height of the 30 mb geopotential (at the center of the cyclone) averages only 2,230 decameters, and the maximum (on the southern periphery of the cyclone) is about 2,380 decameters. The maximum concentration of isohypses, signifying the presence of large horizontal pressure and temperature gradients, occurs near the 60th parallel, where consequently the strongest winds (50-60 m/sec) occur. The low temperature region, outlined by the -75° isotherm, is concentrated near the center of the polar basin. South of the polar cyclone, in the zone between latitudes 40-15° N, as in October, is a region of maximum geopotential (anti--cyclonic region). Over the Pacific Ocean, however, this anti-cyclone occupies a more northerly position and its center is usually located in the region of the Aleutian Islands.

If radiative heat exchange played the deciding role in the formation of stratospheric circulation, as is the case during the summer, then this circulation would obviously be simple and relatively constant. Actually, however, during the cold semi-year the temperature and geopotential fields in the stratosphere experience considerable and varied changes in time and territory, and for this reason atmospheric circulation in the stratosphere during the cold period of the year is rather varied.

Noteworthy are the following principles characterizing evolution of stratospheric processes in the northern hemisphere during the cold period.

1. The geopotential fields are determined by the temperature fields at a /29
given level. Here the regions of minimum geopotential (cyclones) coincide with the regions of relatively low temperatures and, conversely, the fields of maximum geopotential (anticyclones) coincide with the regions of relatively high temperatures. When these fields closely coincide over the polar region zonal circulation is well pronounced on the maps, i.e., westwinds predominate. In these cases a stratospheric cyclone is outlined by several circular isohypses,

which nearly coincide with the parallels. It does not have distinct troughs, or else they occur on its southernmost periphery. This form of circulation conditionally can be called a circular cyclone, a comparatively rare phenomenon.

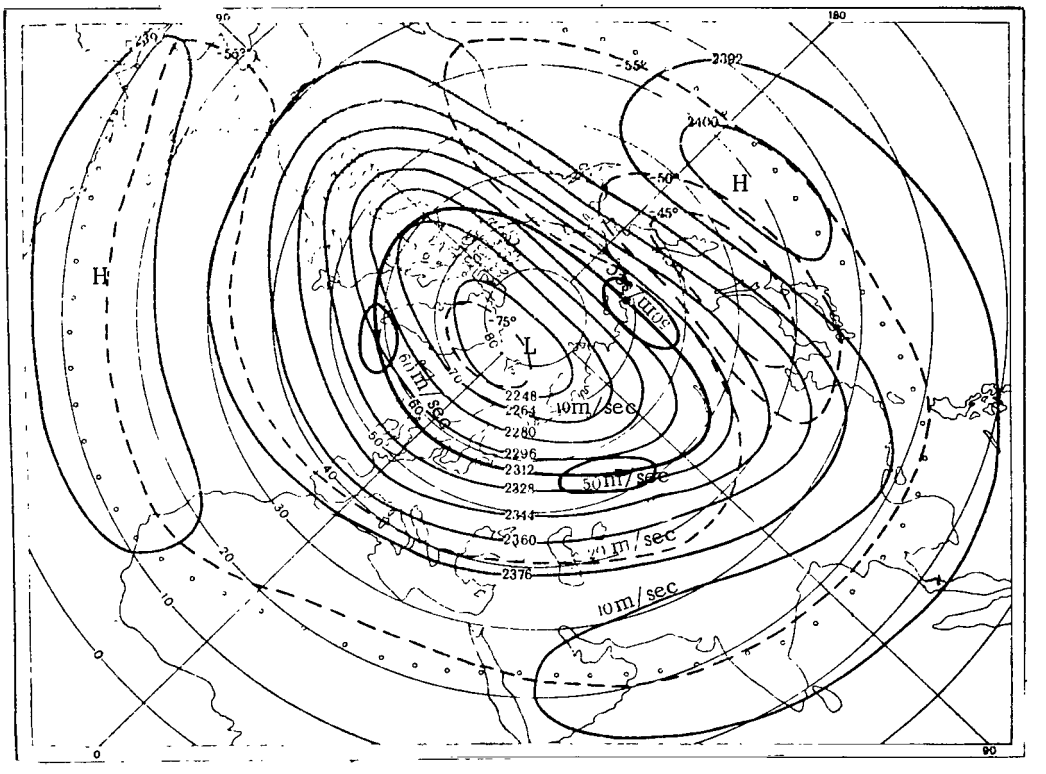


Figure 2.7. Map of Mean Geopotential, Temperature and Wind at the 30 mb Level. January.

2. In the majority of cases cold centers do not coincide with the central part of a cyclone, and the isotherms are generally situated at different angles to one side or the other of the isohypses. It is in this connection that meridional flows occur, which reflect advection of relatively cold air in some regions and of relatively warm air in others. In such cases stratospheric circulation is manifested as a single cyclone, whose center, with one or two, (sometimes more) troughs, extending into the southern latitudes, is located in the arctic region and this means that in the region of a stratospheric cyclone

northwesterly winds prevail in some areas and southwesterly or southerly in others (contiguous) (see Figure 2.8).

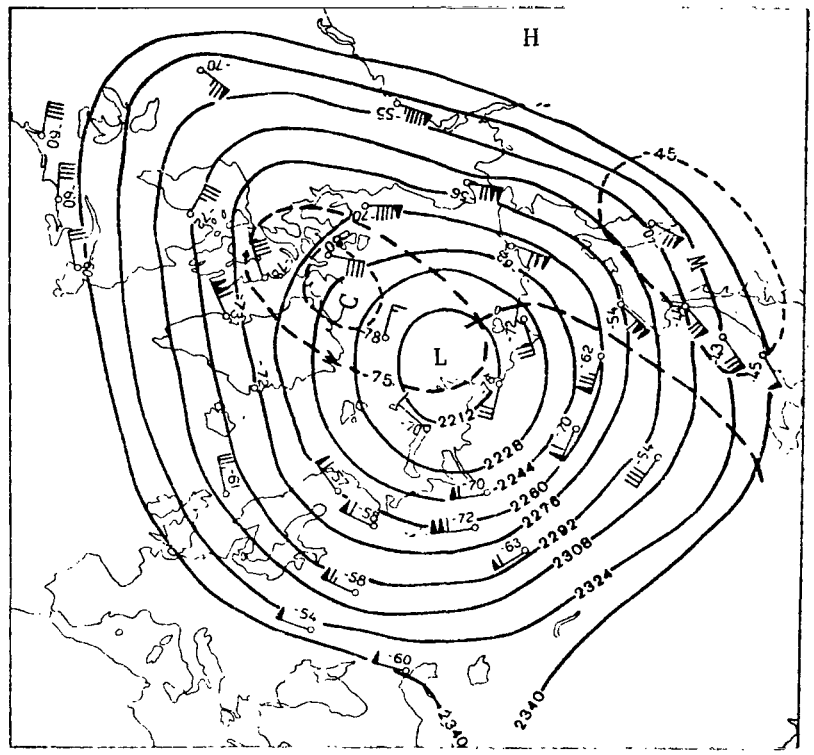


Figure 2.8. AT_{30} Map. 11 December 1967, 1500 Hours.

3. The third type of winter stratospheric circulation is a cyclonic system that includes two independent centers (split cyclone), separated by a ridge, and sometimes by an individual anti-cyclone (Figure 2.9). In this situation there are usually observed appropriately two cold centers, which are of the same intensity for some time, and then change places or reunite into a single center. Sometimes the cyclonic system consists of three or four individual cyclones. Consequently, in certain areas of the northern hemisphere winds of different directions and velocities alternate.

4. The duration of uninterrupted existence of each type of circulation varies. The "circular" cyclonic type of circulation is characterized by the shortest duration (an average of 6.5 days), as are cases of multicentered

cyclones (9 days). Circulations of type 2 (cyclones with a trough) and type 3 (split cyclone) have the greatest duration (an average of 12.2 and 12.0 days, respectively); these varieties of circulation are characterized by a rather long duration (20 or more days) during the first half of the cold season, and a comparatively short duration during the second half. In general, the duration of /30 circulation of types 2 and 3 constitutes between 75 and 88% of all days of the cold season, and the total number of days is divided equally between them.

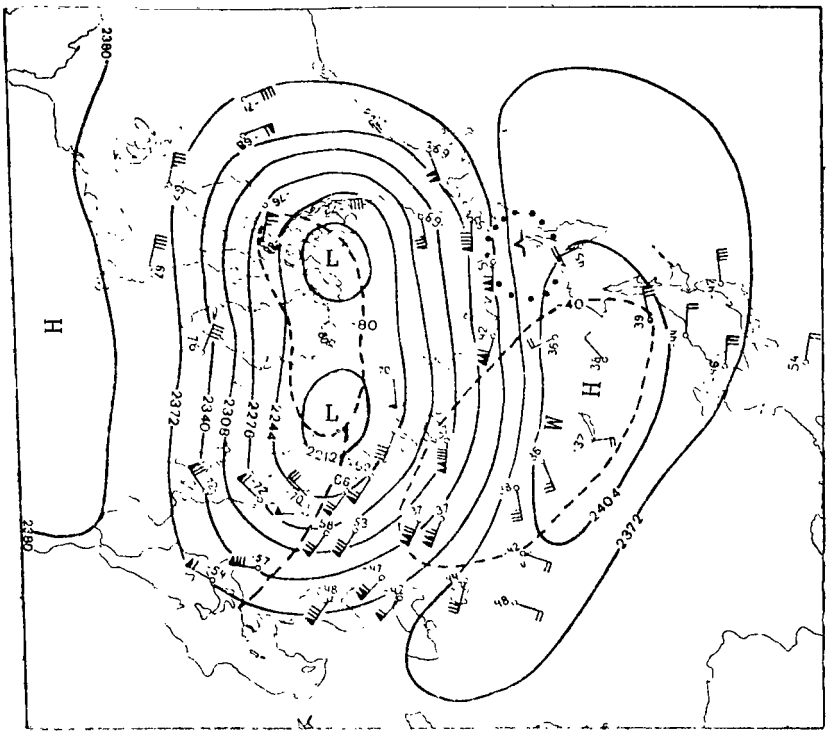


Figure 2.9. AT_{30} Map for 22 December 1967, 1500 Hours.

Changes of the minimum altitudes of the 30 mb level over the central part of the stratospheric cyclone during the period from the first week of October through the 20th-25th of April, are illustrated in Figure 2.10. Comparison of these curves shows that evolution of a stratospheric cyclone does not always correspond to normal seasonal changes in the zenith distance of the sun. In the given case only curve 1, which characterizes in general "normal" evolution

of a winter stratospheric cyclone, comes close to such correspondence. Noteworthy here is that the minimum geopotentials may correspond to periods much later than the hibernal solstice, to the 20th-25th of February in the given case.

Figure 2.10. Variations in Altitudes of 30 mb Level and Center of Stratospheric Cyclones in Different Years. 1, 1966-67; 2, 1967-68; 3, 1968-69.

Curve 2 reflects considerable disruption in the evolution of a stratospheric cyclone, manifested in that during the period between the 15th of December 1967 /31 and the 5th of March 1968 there was a considerable elevation of the 30 mb surface, related to a rather protracted stratospheric potential during that period. Here it is important to note that the indicated potential, as can be seen by analyzing curve 2, was preceded by comparatively sudden temperature reduction and, consequently, by a sharper drop in geopotential than in cases corresponding to curves 1 and 3.

Curve 3 approximates curve 1 through its entire length. Particularly noteworthy for the winter of 1967-68 was the fact that during the beginning of the cold period (approximately until the 10th of December) the stratospheric cyclone developed on the background of substantially reduced geopotentials, and

the stratospheric cyclone evolved on the background of higher geopotentials. This, however, was not related to stratospheric warming, which essentially did not occur during that time. Here, probably, the development of the processes /32
in the troposphere was not completely normal during the winter of interest, when Eurasia experienced continuous Arctic invasion and related processes of long term anti-cyclogenesis. This all indicates how important deviations of geopotential and temperature from certain mean values can be.

Some valuable comments are made on this subject in the work of A. G. Vertlib and G. P. Kalinovskaya [2].

On the basis of statistical processing of data from certain aerological stations located near 50° N latitude and with consideration of both day (1300 hours local time) and night (0100 hours local time) soundings, it is concluded [2] that the annual curve for this latitude should be characterized by a symmetrical temperature change throughout the seasons with the maximum occurring at mid-summer and the minimum at mid-winter. However, as affirmed by the authors [2], several features, most distinct at the transition between the 100 mb and the 20 mb levels, are superimposed on this variation. Thus, the annual curve at the 100 mb level is asymmetric due to the fact that the mean temperatures of January and February exceed somewhat the mean temperatures of November and December. This difference increases at the 50 mb level and the mean temperature for October is below the mean temperatures for January and February. At the 30 mb level the mean temperatures for March and April are lower than for February.

The above enumerated features are characteristic of daytime and nighttime /33
temperatures, but they are more pronounced for the daytime. Hence it follows that at least two mechanisms, specifically the mechanism related to seasonal changes of the zenith distance of the sun and the mechanism that causes winter heating in the stratosphere, most often in January and February, participate in the formation of annual temperature variations in the stratosphere.

2.4. Winter Warming in the Stratosphere

The extremely important and interesting phenomena which we will encounter in our analysis of the stratosphere include cases of considerable temperature

rises, both in the polar region and in the lower latitudes. The first such phenomenon was observed over Berlin between the 26th and 30th of January, 1952, and was investigated by R. Sherkhag. The air temperature at the 10 mb level during the four-day period rose approximately 30°. Analysis indicated that in the given case the temperature first rose at the higher levels and then in the lower layers of the stratosphere. After some time, i.e., in February of the same year, the same phenomenon recurred over Berlin, but this time it was stronger. In this case warming gradually proceeded to the lower layers of the stratosphere. Sherkhag theorized that the principal cause of both stratospheric events were solar eruptions that penetrated into the lower layers of the stratosphere.

In the years that followed Sherkhag proved that the temperature regime of the oceans played a big part in the formation of temperature fields in the stratosphere, and even in the mesosphere. In particular, he assumed that the development of the Aleutian anticyclone as the forerunner of stratospheric heating is greatly influenced by the thermal regime of the oceans.

Since then this phenomenon has been observed and studied by numerous researchers. We now know that the zone of warming encompasses principally the near-polar region and regions of the middle latitudes adjacent to it.

K. Labittske [9] notes that some of these temperature rises are so great as to lead to total destruction of the stratospheric polar eddy, but in all cases this eddy is reestablished after a certain amount of time. A certain relation is seen here between winter temperature rises and the 26-month cycle in the tropical stratosphere.

It has been established that sudden temperature rises are usually preceded by rather sudden cooling. This was also confirmed in our example (see Figure 2.10) at the same time large temperature rises seem to signal the end of winter in the stratosphere--the mean temperature after the completion of such warming remains high. If, however, the fact of temperature rises has become general knowledge, a unique and sufficiently clear theory has not yet been formulated to explain the reasons for its existence. Some authors are in total disagreement with Sherkhag's hypothesis. They cite cases where stratospheric warming

could not possibly be related to solar activity, or when warming does not spread from the upper layers into the lower, but involves simultaneously a considerable thickness of the stratosphere.

G. D Zubyan [6] offered the theory that the chief cause of warming in the /34
stratosphere is interlatitudinal exchange of warm and cold air masses. S. S. Gaygerov [3], who theorizes that temperature rises occur in the Arctic in warm stratospheric air masses "returning" on the eastern periphery of the trough, featuring a high ozone concentration and therefore strongly heated outside of the Arctic region, espouses much the same hypothesis. The following comments should be made concerning this explanation. Actually, when a baric trough is formed somewhere in the stratosphere, a temperature contrast occurs simultaneously, particularly notable along the axis of this trough. Rather strong south winds and relatively high stratospheric temperature are observed east of the axis. West of the axis of the trough, air at a lower temperature moves from north to south. The temperature contrast along the trough axis can reach the order of 10° for 100 km of distance. At first glance this is analogous to the situation observed during analysis of tropospheric fronts on the earth's surface. Actually, however, there are important differences in the formation and evolution of stratospheric troughs compared to tropospheric troughs.

a. The movement of stratospheric troughs is in no way related either to wind direction or to wind velocity, as is the case in the lower layers of the troposphere. As a rule, the axis of a stratospheric trough may remain for several days over the very same region or move only slightly in some direction, and then shift suddenly several hundred kilometers to the west or east, usually in the direction where a new cold center has formed.

Sometimes, in the general southwesterly flow of relatively warm air, high temperature centers (up to 30-35° at AT_{30}) form in the leading part of the stratospheric trough; despite the high wind velocities, these centers remain over the same location for several days, and then move hundreds of kilometers to the west in just 12 to 24 hours. This is a rather frequent phenomenon along the east coast of Asia and once again it can be explained by development of a

new warm center to the west or south of the previous one. Consequently, stratospheric warming cannot be attributed simply to advection of warm air.

Kh. P. Pogosyan and A. A. Pavlovskaya [12], and also in recent years Kh. Kh. Rafailova [17], established that during winter heating in the stratosphere over the Arctic the cyclonic eddy splits and the anticyclone retreats to the middle and high latitudes. Such splitting is preceded by two-way advection of heat in the troposphere from the Atlantic and Pacific Oceans to the polar region. /35

A. L. Kats [8] propounds that stratospheric temperature rises in the northern hemisphere are fluctuational invasions of warmth into the winter stratosphere from the warm anticyclonic stratosphere of the Southern Hemisphere; these invasions explain the formation or intensification of tropical anticyclones and the meridional formations of the entire baric field; due to the lack of observation data this hypothesis cannot be checked against factual material.

We could cite many other works in which cases of substantial warming in the stratosphere are explained one way or another. Worthy of greatest attention, in our opinion, are the concepts of the causes of winter warmings expounded by M. V. Shabel'nikova [19]. The basic theories and conclusions of her work consist in the following.

1. Circulation in the stratosphere, leading to stratospheric temperature rises, is a result of meridional transformations in the troposphere; the process of meridional rearrangement of the baric and thermic fields begins at the earth's surface and spreads rapidly through the entire troposphere, and then into the stratosphere.

2. As a result of meridional rearrangement in the troposphere, an upper-air trough is formed over a certain region, in which system prevail ascending vertical motions; in the stratosphere, however, at some altitude over the given region, the sign of the vertical motions changes, i.e., descending motions occur. These motions also lead to heating of the air at the corresponding altitudes and to the formation of a warm center in the stratosphere; the adiabatic temperature rise in the given case usually considerably exceeds its advective drop.

Correspondingly, during formation in the troposphere of an upper air ridge, ascending motions develop above it in the stratosphere, leading to the formation here of a cold center. As a result, a trough forms in the region of the winter stratospheric eddy, along the axis of which are formed more or less substantial temperature contrasts: to the left (to the south) of the trough axis moves relatively cold air, and to the right (to the east), relatively warm air.

3. The temperature rise in the various layers of the stratosphere differs for the very same magnitudes of descending motions: the smaller the vertical temperature gradient the higher the air temperature rises. In the troposphere, /36 where the air temperature decreases with altitude, descending motions lead to a smaller temperature rise than in the lower stratosphere, where isothermia takes place. In the stratosphere, at altitudes from 25-30 to 50 km where there is constant inversion, the temperature rise will be much greater than in the lower stratosphere. Thus, for instance, when the velocity of the descending motions is 3 cm/sec in the troposphere for $\gamma = 0.6°/100$ m, the temperature will rise 10° during the day, and in the inversion layer of the stratosphere for $\gamma = -0.4°/100$ m, the temperature for the very same vertical velocity will rise 36°.

Thus, above 25-30 km even slight descending motions lead to a sharp temperature rise. Beneath the strengthening warm region the inversion layer descends and conditions are created for more rapid adiabatic temperature rise and spreading of warmth from the upper layers into the lower.

It should also be pointed out that processes which give rise to stratospheric warming may develop on various scales. In most cases these warming processes are limited only to the leading part of the stratospheric trough. During certain years, however, they become so large that they lead to enormous disruptions of winter stratospheric circulation, manifested in the formation of individual anticyclones in the high and middle latitudes. In order to understand such processes we will examine a case of considerable warming in the stratosphere of the northern hemisphere in December, 1967 and in January 1968.

On December 11, 1967, the stratospheric cyclone represented a single center, located in the region of Severnaya Zemlya (See Figure 2.8) with a distinct trough, directed toward Hudson Bay, and a less distinct trough, oriented toward the middle course of the Amur River. In the given case we are interested in the second trough. West of the axis of the trough (broken line) there is a westward flow of relatively cold air at temperatures between -60 and -72°; east of the axis is southwestward flow at temperatures between -40 and -52°. The center of warmest air (about -40°) was located over Kamchatka; the coldest air center (about -80°) was located in the region of Baffin Island.

The center of the Aleutian anticyclone was located approximately 1,000 km south of Alaska. The center of the anticyclone over the Atlantic Ocean was approximately 1,500-2,000 km from the Azores Islands.

During the ensuing twelve days the center of the stratospheric cyclone moved to the west, i.e., from the region of Severnaya Zemlya to the region of Medvezhiy Island (Barents Sea); by the 22nd of December it split, whereupon a second center formed a little to the west of Baffin Bay (Figure 2.9). The axis of the trough of the main cyclone over the Barents Sea then passed from north to south, approximately along the western boundaries of the USSR, where considerable horizontal temperature contrasts and quite strong southwesterly winds /37 were observed. To the west of the trough axis spread relatively cold stratospheric air with temperatures at the 30 mb level of -70 to -76°. The temperature contrasts in the region of the trough between Moscow and Kaliningrad at the 50, 30 and 20 mb levels were 2.8, 11.0 and 28.0°, respectively. It was important to determine the features of the structure of the troposphere and stratosphere in the examined region and to what extent the above-described hypothesis of M. V. Shabel'nikova concerning conditions of the development of stratospheric heating, is valid. For this purpose a vertical cross-section of the atmosphere was constructed (Figure 2.11) for the layer from 1,000 mb to the 20 mb level in the /38 direction from London to Kurgan. Shown on the cross-section are the tropopause, isotherms, direction and velocity of the wind. What stands out most here is the ridge of the tropopause with the peak above Hannover, to which a warm barometric ridge corresponds in the troposphere. It is thought that this tropospheric ridge

is characterized by descending vertical movements. Simultaneously in the stratosphere, in approximately the 100-20 mb layer, is a cold air center with a temperature of -75 to -77° and, obviously, this center occurs in the region of ascending vertical movements. The wind in the examined region was northwesterly at 30-40 m/sec.

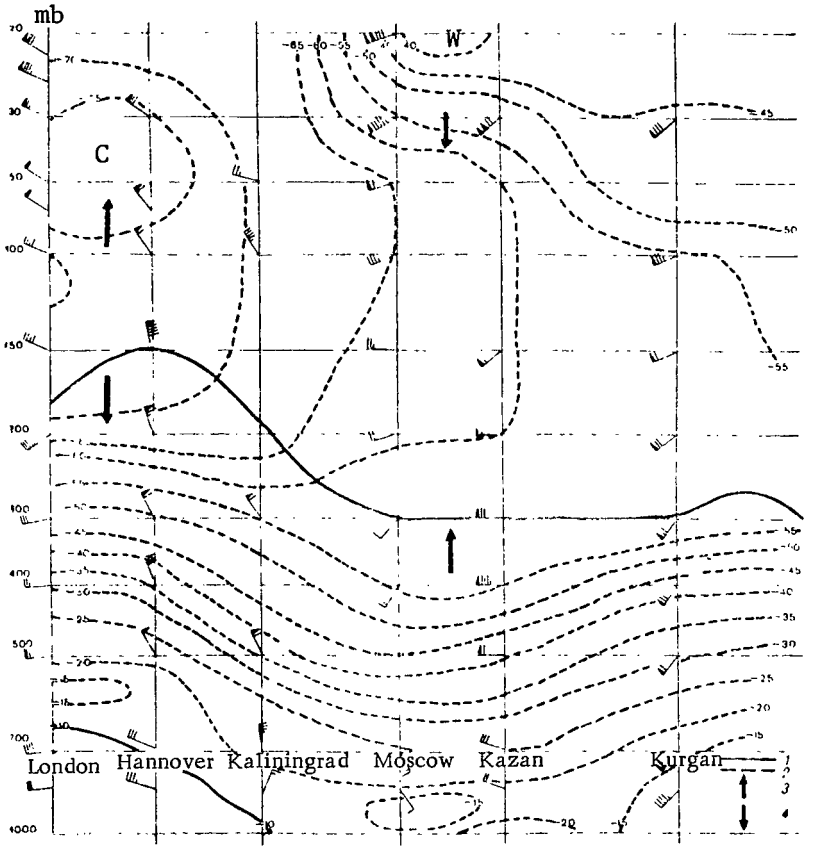


Figure 2.11. Vertical Cross-Section of Atmosphere in London-Kurgan Direction. 22 December 1967, 1500 hours. 1, Tropopause; 2, Isotherms; 3, Ascending vertical movements; 4, Descending.

In the eastern part of the region we see quite the opposite picture. Specifically, the tropopause is considerably lower due to the presence in the troposphere of a thermal, and consequently baric trough, characterized by

ascending vertical movements. Above the tropospheric trough, in the middle stratosphere, is a distinct warm center, characterized by vertical descending movements, in which region strong southwesterly winds prevail.

Thus the theory advanced in [19] to the effect that more or less substantial temperature changes in the stratosphere are related to vertical air movements, is verified. Particularly, considerable warming occurs above the 50 mb level.

Then the Aleutian anticyclones seemed to shift from east to west and its influence spread over nearly all the territory of the USSR, causing considerable stratospheric warming. Center (W) of maximum temperature rise (to -34°) was in the region of the middle course of the Yenisey River.

In connection with the splitting of the polar stratospheric cyclone (December 22), the processes over North America became distinctly meridional, with the result that the temperature in the leading part of the trough began gradually to rise and relatively warm air began to spread to Greenland, northern regions of the Atlantic and northwestern and western regions of Europe. In this connection the former trough to the west of the European part of the USSR moved rapidly to the eastern regions of the USSR and an independent anticyclone formed over western Europe (Figure 2.12).

During the ensuing period splitting of the stratospheric cyclone further intensified and the stratospheric anticyclone process between the two cyclones strengthened (Figure 2.13) on the background of relatively high temperatures in the region of Baffin Bay and Greenland Sea (to -34, -39°). The process of anticyclogenesis continued to intensify until January 13 and was reflected in the existence of two independent anticyclones, one over Greenland and the other /39
over the Chukot Peninsula. The more active of these was the anticyclone over Greenland.

It is often said that winter stratospheric warmings are of an explosive character, i.e., are distinguished by an extremely rapid temperature rise. Thus, in the case recorded by Sherkhag in January 1952, the temperature at the 10 mb level rose 30° in four days, i.e., an average of 7.5° per day. There have been cases when the temperature rose 18° in 12 hours. In this connection it was

important to find out how the temperature in the stratosphere changed in the described case of warming in December, 1967. For this purpose, graphs of temperature change from one day to the next were constructed for five points, ranged from west to east--from London to Khabarovsk (Figure 2.14). These graphs are not highly accurate because of frequent omissions of observations, particularly for the 10 mb level. In the final analysis, however, they are quite important. /40 Temperature changes at the 50, 30 and 10 mb levels were examined.

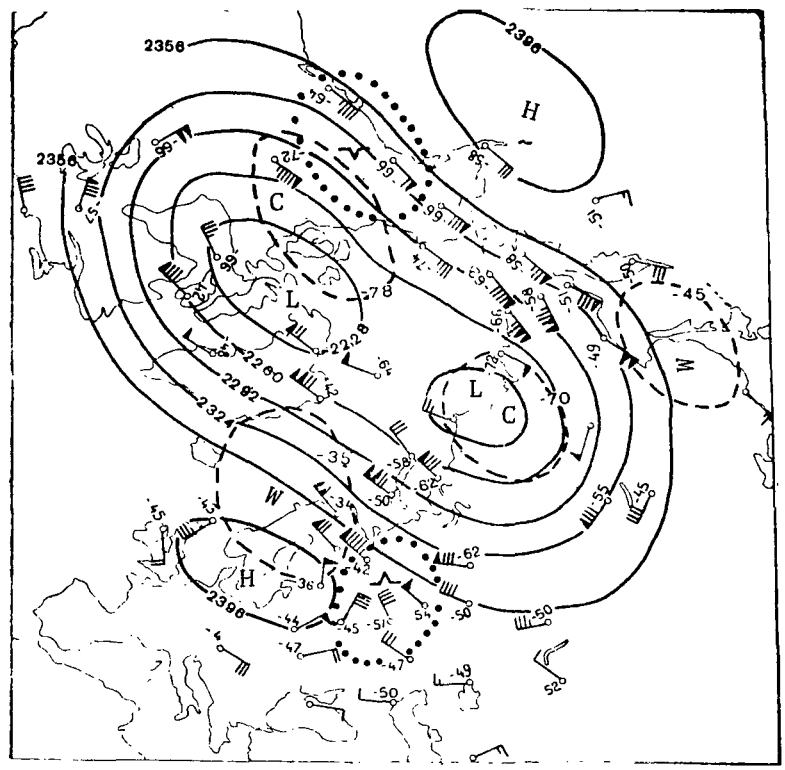


Figure 2.12. AT_{30} Chart. 29 December 1967, 1500 hours.

We will discuss the following features of temperature changes both in time and in space for the period of the examined stratospheric warming.

1. The greatest relative temperature changes occurred over Europe and western Siberia, and the smallest over the Far East where a high temperature background, in the limits of -39, -47°, lasted for the entire month. Considering the definite relationship that exists between processes in the troposphere and stratosphere, it can be assumed that maximum temperature fluctuations in the

stratosphere over Europe and western Siberia (Region II) occur in connection with the great variability of tropospheric processes above these regions in comparison with the variability of these processes in the Far East. /41

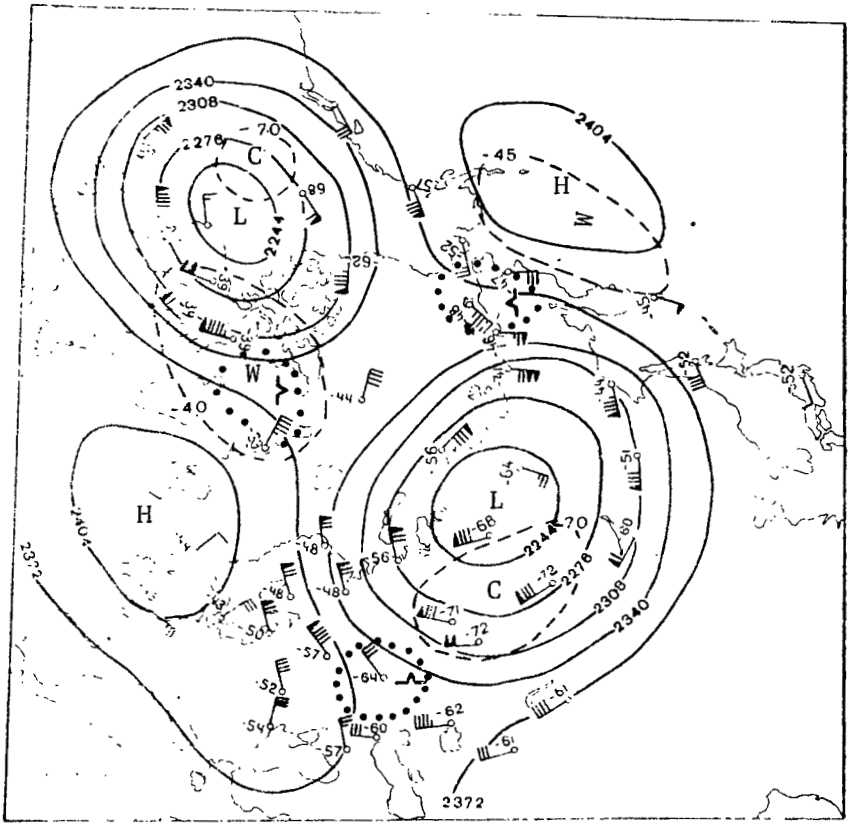


Figure 2.13. AT_{30} Chart. 6 January 1968, 1500 hours.

2. The temperature curves at all three levels are almost identical for all five points, i.e., the temperature rise (dropping) at one level corresponds to the analogous temperature change on the other levels.

3. The temperatures at the 50 and 30 mb levels differ little; in most cases the temperature at the 50 mb level is 1 to 5° lower than at the 30 mb level, but /42 in individual cases the temperature difference at these levels is 10-12°.

4. Much greater temperature differences exist between the 30 and 10 mb levels, particularly in cases of an over-all temperature rise in the stratosphere, when these differences reach 20-25°.

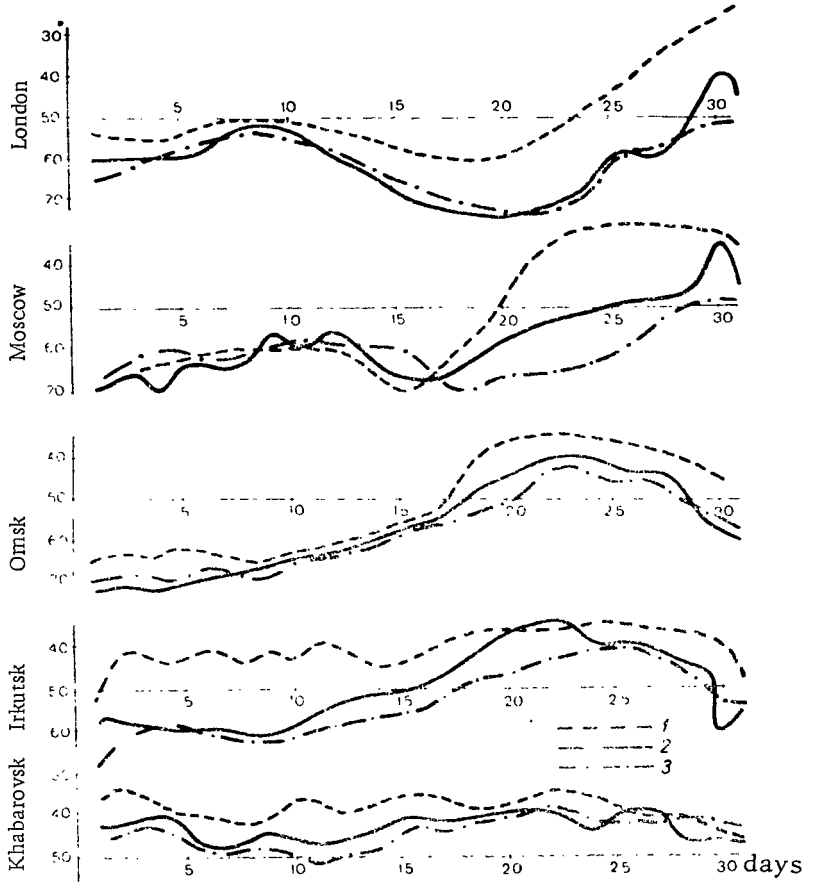


Figure 2.14. Curves of Temperature Change at AT_{30} Level for December, 1967. 1, 10 mb; 2, 30 mb; 3, 50 mb.

5. The temperature changes in time are of a wavelike character, but the wavelength can differ considerably. The temperature rise phase, judging by data from London, Moscow, Omsk and Irkutsk, lasts for about 12 days.

2.5. Jet Streams in the Stratosphere

Jet streams, as we know, are comparatively narrow zones of strong winds comprising in cross-section several hundred kilometers with a length of several thousand kilometers. At the core of the jet stream, known as the axis of the stream, the wind velocities are maximum and reach, as a rule, 60 to 80 m/sec. To the left and right of the axis, looking in the direction of the wind, at distances of 200-400 km the wind velocities are much lower, particularly on the left. In the vertical direction the jet streams are also characterized by a sharp strengthening of the wind in the lower portion of the stream and by the corresponding weakening in the top part. In other words, jet streams are bounded on the left and right, bottom and top, by high horizontal and vertical wind velocity gradients, known as wind shears.

The position of the jet streams always coincides with the zones of maximum horizontal temperature gradients, i.e., for those regions in which the warm and cold air masses are closest together. In other words, the jet streams are related to high altitude frontal zones.

Tropospheric jet streams have been analyzed quite thoroughly, but stratospheric jet streams have not. Nevertheless, the above-mentioned characteristics are obviously general, both for tropospheric and for stratospheric jet streams.

We will examine the problem of the conditions under which jet streams develop in the stratosphere. During the warm half of the year these streams exist only in the equatorial zone of low pressure, along the southern periphery of the summer stratospheric anticyclone. They appear as easterly winds, blowing at a velocity of about 30 m/sec. The maximum wind velocities occur in July, when the stratospheric anticyclone is in a stage of maximum development. Here we will examine jet streams in the middle and partially in the northern latitudes, occurring during the cold part of the year. The first impression of the jet streams can be obtained by examining AT_{30} charts, which illustrate the synoptic conditions of the development of stratospheric warming (see Figures 2.12-2.15). Analysis of these maps lead to the conclusion that jet streams /43

above Eurasia can flow in different directions. Usually, however, they appear as strong winds of the westerly or southwesterly directions.

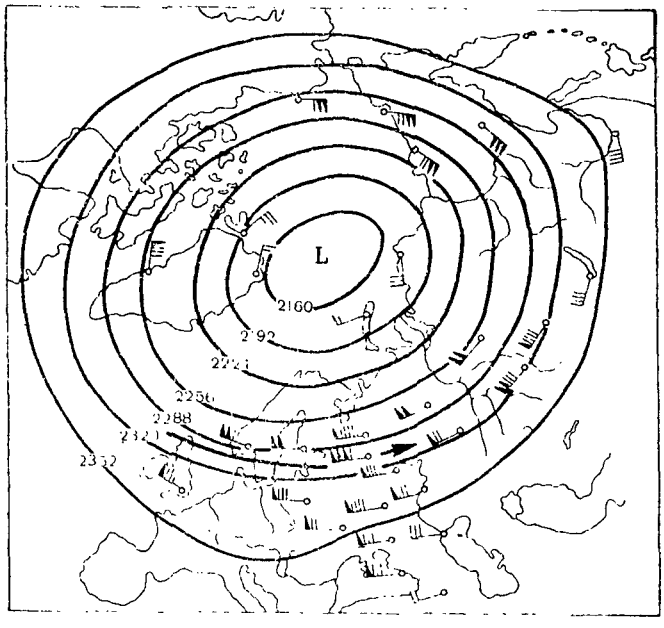


Figure 2.15. AT_{30} Chart. February 25, 1967. 1500 hours.

Since the jet streams occur in regions where considerable horizontal temperature gradients exist, there may be several reasons why considerably horizontal temperature contrasts, and consequently, strong winds, occur in the stratosphere. One of them is considerable radiative cooling of the stratosphere during the polar night over the Arctic, i.e., at latitudes from the Polar Circle (66°31') to the Pole. Arctic air in such cases is much colder than the air of the middle latitudes. Thus a frontal zone occurs, characterized by rather strong west winds. A good example of such a situation is the jet stream of 25 February 1967.

The AT_{30} chart compiled on the basis of observations for 15 hours on the 25th of February, 1967, is shown in Figure 2.15. It indicates the existence on this day of a stratospheric cyclone centered almost over the Pole itself. The isohypses outlining this cyclone are directed almost strictly in the latitudinal direction, from west to east. This is a typical zonal jet stream, whose axis passes approximately along the latitude 60° at a wind velocity of about 85 m/sec. The thinning out of the isohypses towards the north from 70° North latitude indicates the existence of a considerable cyclonic wind fault. /44

Such thinning out of isohypses also occurs at the southern extremity of the jet stream zone, approximately along the 50th parallel; this is an anti-cyclonic wind shear.

The width of the entire jet stream zone is approximately 1,000 km.

Figure 2.16 shows the vertical cross-section of the examined jet stream in the direction from Vis Island to Yerevan. Unfortunately, due to the lack of necessary observations, the Arctic Region has been studied very little, and the vertical cross-section presented herein does not strictly correspond to the meridional direction.

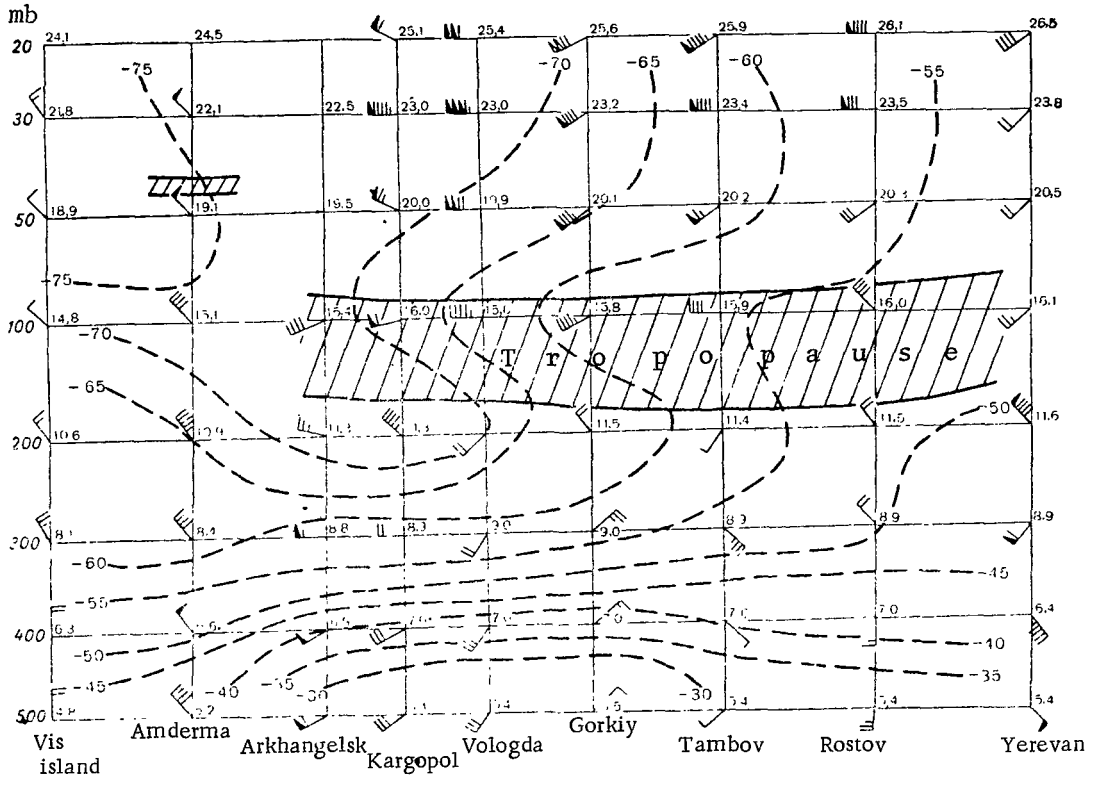


Figure 2.16. Vertical Cross-Section of Atmosphere in the Direction From Vis Island to Yerevan. 25 February 1967, 1500 hours.

Noteworthy are the following features of the temperature distribution and wind for this case.

1. In the Arctic (Vis Island--Amderma) the tropopause occupies a rather high position (19-22 km), which does not occur at the middle latitudes. Such a high position of the tropopause can be considered normal for the Arctic Region, where there is a prolonged radiative cooling of the air. From Vologda to Yerevan there is a rather thick inversion layer with a lower boundary at an altitude of about 12 km and the upper boundary at an altitude of about 16 km, which can be taken as the tropopause of the middle latitudes.

2. The lower boundary of the stratospheric jet stream occupies the lowest position (15-16 km) on its cyclonic side, in the region of Kargopol-Vologda in the given case. To the south, toward the anticyclonic side, the lower boundary of the jet stream rises (20.1 km at Gorkiy, 23.5 km at Rostov). The maximum wind velocities (exceeding 80 m/sec) occur approximately along the 60th parallel.

3. By comparing the given vertical cross-section with the analogous cross--section shown in Figure 2.11, we discover first of all, good similarity between them in the horizontal temperature distribution and wind distribution in the stratosphere and, secondly, quite the opposite picture in the vertical temperature distribution in the troposphere. As stated above, stratospheric heating can occur in the case when ascending movements, to which correspond descending movements in the troposphere and vice versa, occur in the troposphere. Actually, in the given case the appearance of considerable horizontal temperature contrast in the stratosphere was related not so much to vertical movements in the atmosphere as to the different rates of radiative cooling of the air in the Arctic basin and at the middle latitudes.

A strictly zonal jet stream, directed along the latitude circles, is a /46
comparatively rare phenomenon. In the vast majority of cases zonal circulation has a meridional component, appearing for a number of different reasons.

In this connection it can be said that the following play a tremendous role in the development of jet streams: first, different rates of radiative cooling of the stratosphere on the polar and middle latitudes; secondly, vertical movements in the troposphere and stratosphere; thirdly, stratospheric

warming and, fourthly, action in the northern direction of anticyclonic formations occurring in the stratosphere at the middle latitudes.

By analyzing these factors we conclude that they are all so interrelated that it is absolutely impossible to distinguish the principal and secondary factors among them. They seem to act simultaneously, and specifically when atmospheric circulation changes in some region from zonal to meridional and when opposite vertical movements occur in this connection.

As stated above, in one such region, ascending movements occur, resulting in cooling of the air in the stratosphere, and in another--descending movements, leading to heating of the air. As the result, more or less considerable horizontal temperature gradients and, accordingly, strong winds develop.

High geopotentials are created in the region of relatively warm air, involving the formation of stratospheric cyclones which, being in the vicinity of the basic stratospheric cyclones, create large horizontal pressure gradients and, consequently, strong winds. The stratospheric cyclone in such cases is extended in length or width, with strong southerly winds from one side and northerly winds on the other, or with westerly winds on one side and easterly on the other.

We will discuss briefly the relationship between tropospheric and stratospheric jet streams and between changes in wind velocity with altitude and the corresponding temperature changes. We will illustrate this by analyzing three typical cases.

Stratification and change of the wind with altitude in the troposphere and stratosphere (starting at the 500 mb level) are shown in Figure 2.17 for the most typical cases on the basis of the data from atmospheric radio sounding in Ashkhabad, Murmansk and Moscow.

In Figure 2.17a we see a typical tropospheric jet stream with a maximum wind velocity of 48 m/sec, 1.5 km below the tropopause; above the tropopause the temperature curve is almost isothermic to the maximum elevation (about 29 km). The wind weakens in the entire layer of the stratosphere, starting especially at the 100 mb level).

In Figure 2.17b, the situation in the tropopause and lower stratosphere to the 70 mb level is the same in principle as in the preceding case (Figure 2.17a) /47 The tropospheric jet stream has about the same intensity with maximum wind velocity of about 45 m/sec beneath the tropopause. Above the tropopause, to an altitude of about 18 km, the temperature rises slowly and the wind velocity drops slowly. Starting at 18 km the temperature begins to rise more rapidly, and the wind strengthens considerably.

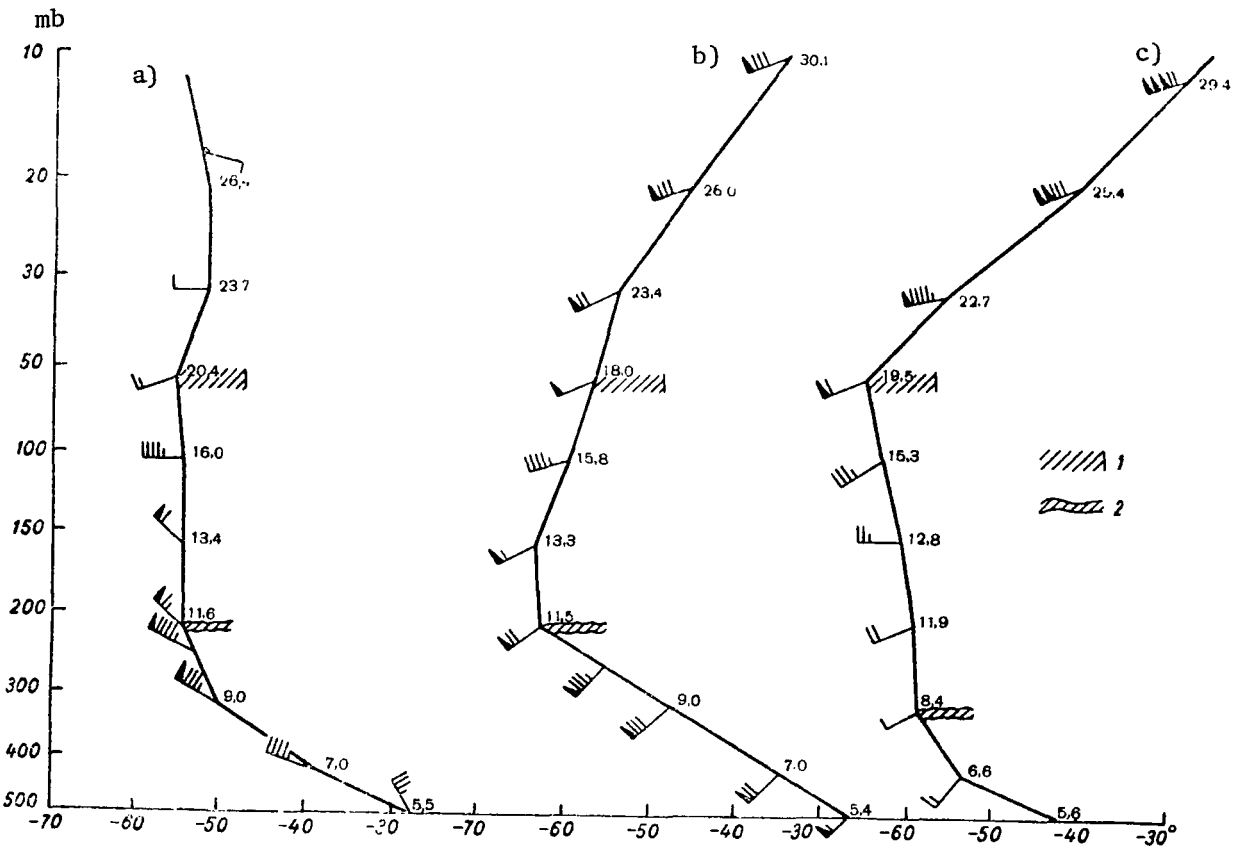


Figure 2.17. Characteristic Changes in Wind Velocity with Height in Connection with the Features of Change in Stratification. a, Ashkabad, 15 hours, February 17, 1969; b, Murmansk, 1500 hours, February 17, 1969; c, Moscow, 1500 hours, December 21, 1967; 1, Isopause; 2, Tropopause.

The third case (Figure 2.17c) is characterized by the presence of weak wind, in the 5-15 m/sec range in the troposphere and in the stratosphere to the 50 mb level (19.5 km). From 19.5 km the temperature began to rise sharply and the wind began to increase and reached a velocity of 85 m/sec at an altitude of 29.4 km.

On the basis of the above it can be concluded that there is a certain relationship between temperature changes with altitude and the corresponding wind velocity changes, specifically: in the troposphere the increase in wind velocity with altitude is related to the corresponding temperature drop; in the stratosphere, on the other hand, the increase in wind velocity with altitude is related, as a rule, to distinct inversions, i.e., to an increase in temperature with altitude.

Figure 2.17 clearly shows the level of the isopause. It is located approximately at the 50 mb level, and cases of sharply pronounced inversion are characterized by considerable strengthening of the wind.

2.6. Interaction of Processes in Troposphere and Stratosphere

The question of interaction of processes in troposphere and stratosphere becomes more important with each passing day. The study of this problem is essential in the interest of the development of methods of short-term and long--term general weather forecasting and of methods of forecasting weather for aviation.

Certain principles have been established that indicate the relationship of the development of processes in the troposphere and stratosphere for the winter period of the year. One good example of such interaction is the above-described process of the development of a baric trough in the stratosphere, one side of which forms relatively cold air and the other side, relatively warm air. As a consequence of this considerable horizontal temperature contrasts develop along the axis of the stratospheric trough, which form as a result of vertical movements with the opposite signs compared to the vertical movements in the troposphere over the given region.

In the earlier investigations of Kh. Pogosyan, A. A. Pavlovskaya and M. V. Shabel'nikova [11], the question of how changes of the baric field in the troposphere cause corresponding changes in the stratosphere was examined rather thoroughly. They conclude, in particular, that the splitting of the stratospheric cyclone is a result not only of cooling of the air above a continent, but also of intensive processes of cyclogenesis in the troposphere, developing from Greenland in the direction of the Kara Sea. /49

To this may be added also the following features of the development of processes in the troposphere in the specified region, revealed by us through analysis of a number of cases. Toward the moment of splitting of the stratospheric cyclone a deformation field, one of the components of which is a deep cold cyclone centered in the region of the Taymyr Peninsula, and the other a high warm anticyclone centered approximately in the region of Vologda, is formed in the troposphere over the northern regions of Europe and partially over western Siberia. As a result of this an active high altitude frontal zone, extending from the Barents Sea to the southern part of the Ural Mountains, is formed.

The analogous high altitude frontal zone occurs over the North Atlantic, extending from Greenland to the Barents Sea. Thus, favorable conditions are created for the development of a series of cyclones, followed by the concluding invasion of cold air from the direction of the Barents Sea toward the European Continent. During this time, as we know, a rather powerful anticyclone develops, strengthening to the stage of conversion to a high altitude warm formation. The latter, together with the adjacent high altitude cold cyclone, forms a new high altitude frontal zone, and the entire cycle of processes is repeated. This cycle of processes usually lasts for six to seven days, i.e., within the limits of duration of the so-called natural synoptic period, during which the stratospheric cyclone is in the split state. If, however, monotypical periods are repeated several times, the duration of existence of a dual center stratospheric cyclone increases correspondingly (sometimes up to 28 days).

It was stated above (2.4) that winter stratospheric circulation is manifested as three basic forms, specifically "circular" cyclone, cyclone with one or

two troughs and split cyclones. In order to determine more objectively to what extent a relationship exists between the established principles of evolution of a stratospheric cyclone, on the one hand, and changes in the development of tropospheric processes on the other, the circulation indices proposed by A. L. Katz [7], in pertaining to the characteristics of circulation in the troposphere at the AT_{500} level, were used. Here the general index of circulation I', characterizing the relation between the meridional I_m and zonal I_z indices $I' = I_m/I_z$, was examined. The index was calculated for three regions of the Northern Hemisphere:

Region I: Eastern Atlantic--Europe--Western Siberia;

Region II: Eastern Siberia--Western part of Pacific Ocean:

Region III: Eastern part of the Pacific Ocean--North America, Western Atlantic. /50

Analysis of the calculations revealed that the transition from one form of stratospheric circulation to another is always accompanied by a change to one or the other side of the general index of tropospheric circulation. The greatest changes occur over Region I, the least over Region II. Generally speaking, however, in the majority of cases I' = 1.0 for all three regions, i.e., the zonal component of circulation clearly predominates. This predominance is most clearly pronounced in Region II; above this region even the direction of development of the processes is most stable in time. The most contrasting values of I' correspond to circulation of forms 2 and 3. In these cases the meridional component of circulation in the troposphere can exceed considerably the zonal component. In the first type of stratospheric circulation (circular cyclone) the zonal component predominates in all cases.

Along with the use of circulation indices, determination of the relations between processes in the stratosphere and in the troposphere was accomplished by constructing composite maps of the trajectories of the movement of near-earth centers of cyclones and anticyclones during the periods of existence of certain forms of cyclonic circulation in the stratosphere. Comparison of such maps revealed that a certain direction of development of synoptic processes in the troposphere corresponded to each form of stratospheric cyclonic circulation.

We will examine two adjacent periods by way of example.

The period beginning October 23 and ending November 3, 1966 (lasting nine days). Cyclonic circulation in the stratosphere was determined by the existence of a single cyclone with its center near the Pole (form 1). The character of processes in the troposphere can be judged upon the basis of the trajectories of movement of near-earth centers of cyclones and anticyclones (Figure 2.18), which indicate clear predominance of the zonal component of air mass transfer in the troposphere.

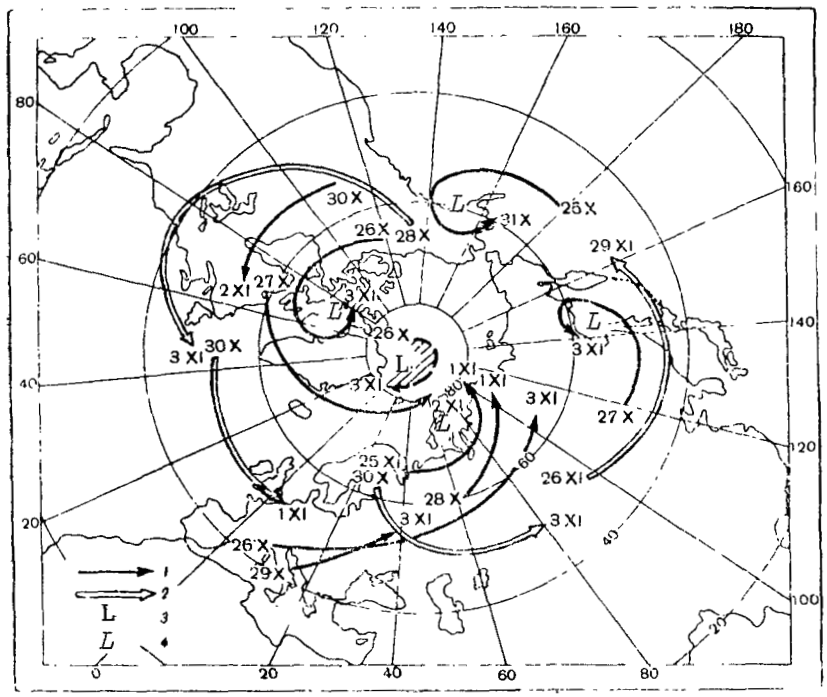


Figure 2.18. Composite Kinematic Map for Period Beginning October 23 and Ending November 3, 1966. 1, Trajectories of cyclones; 2, Trajectories of anticyclones; 3, Centers of stratospheric cyclones; 4, Centers of filling tropospheric cyclones. The regions where the centers of stratospheric cyclones are located are cross-hatched.

We will note the following features of the development of processes in the troposphere during a given period.

1. These processes included both series of cyclones and conclusive invasions of anticyclones, of which the most notable were invasions during the second half of the period from Ladoga Lake to the southeast of the European territory of the USSR and further toward Lake Balkhash, and also invasion over North America, approximately from Alaska to Lake Michigan and further to the east-northeast. By the end of the period these anticyclones reached their maximum intensity.

2. Cyclones of individual series became concentrated toward the end of the period in certain regions, specifically in the regions of Hudson Bay, Kara Sea, Okhotsk Sea and Gulf of Alaska. As a whole, however, the entire system of cyclones and anticyclones more or less rotated counterclockwise without deviating substantially from the latitudinal direction. /51

Period beginning November 4 and ending November 11, 1966 (duration 8 days). The stratospheric cyclone of the preceding period split on November 4, so that the center of one of them was located over northern Greenland, in the center of the other over the Taimyr Peninsula. Cyclone splitting coincided with accentuation of the frontal zone in the troposphere, passing from the Barents Sea to the middle course of the Ob and Yenisey Rivers.

Comparison of the map (Figure 2.19) for the period of November 4 through 11 with the map for the preceding period reveals considerable changes in the directionality of development of the processes in the new period. These changes are manifested primarily in an increase of meridionality of circulation in the troposphere over the continents of Europe and North America.

Circulation in the stratosphere also appeared to separate. Here, if in the preceding period the trajectories of cyclones over Europe were directed from the southwest to the northeast, in the new period they were directed in most cases from the northwest to the southeast. As in the preceding period, the tropospheric processes are expressed first in the passage of a series of cyclones, and then in the concluding invasion of anticyclones from the Arctic Basin. Invasions of such anticyclones in the intensification stage characterize /52

the essence of atmospheric processes in the second half of the examined period. It should be pointed out here that the other anticyclonic trajectories shown in Figure 2.19 pertain either to the preceding anticyclones of Arctic invasion, already in the stage of gradual disintegration, or to anticyclones formed in the air masses of the middle latitudes.

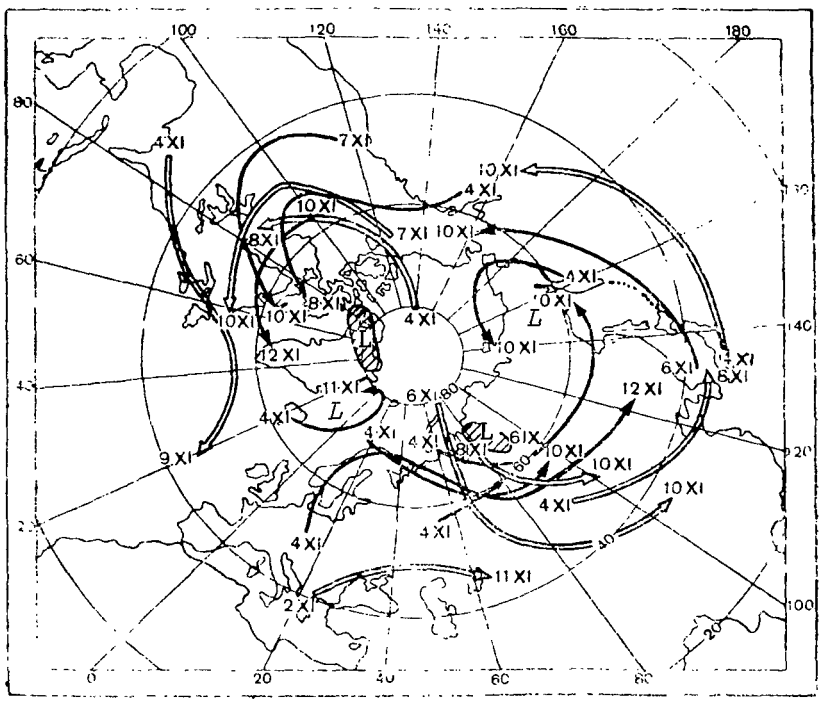


Figure 2.19. Composite Kinematic Map for the Period from November 4 to 11, 1966. For meaning of notations, see Figure 2.18.

The principles of evolution of the winter stratospheric cyclone presented herein, in connection with the directionality of development of processes in /53 the troposphere, enabled meteorologists to acquire a more accurate orientation for compilation of stratospheric wind and temperature forecasts. The transition from one form of circulation to another is usually accompanied by substantial rearrangement of the wind and temperature fields. The new type of processes should last for several days, during which period there are usually no sudden changes in the wind and temperature regime.

As concerns the warm period of the year, when the circumpolar anticyclone is formed, there is no direct relationship, as indicated by analysis of daily synoptic maps for the various levels of the troposphere and stratosphere, between the processes of the two spheres. Meanwhile we know that certain deviations from the normal course of evolution of the stratospheric anticyclone occur from year to year (see 2.3). S. T. Pagava [10] offers some interesting data, indicating the existence of a certain relationship between the periods of formation of the stratospheric anticyclone over the Pole and the ensuing thermal regime in the troposphere over the European USSR and western Siberia. S. T. Pagava, in particular, established that the cold spring of 1964 was related to the early formation of some recirculation in the stratosphere, and the warm spring of 1962, to the late formation of the stratospheric anticyclone.

Research on the processes in the stratosphere and their relationship to processes in the troposphere, conducted in recent years, are of tremendous importance for the development and perfection of weather forecasting methods. There is little doubt that the information that we will obtain from aviators flying in the stratosphere will be extremely useful in the solution of theoretical and practical problems of the weather service.

CHAPTER 3

PRINCIPLES OF VERTICAL DISTRIBUTION OF AIR TEMPERATURE AND WIND TO ALTITUDE OF 30 km AND THEIR SIGNIFICANCE FOR SUPERSONIC FLIGHT /54

3.1. Seasonal Features of Vertical Temperature Distribution

The vertical distribution of air temperature is important in the determination of engine operating mode during ascent of a supersonic aircraft to cruise altitude, particularly in transition from subsonic to supersonic velocity above 8-12 km.

During the various seasons of the year both temperature and its vertical gradients differ substantially in the troposphere and the lower stratosphere. The curves of vertical temperature distribution are shown in Figure 3.1 for the Polar Region and the 80, 60, 50, 40, 30 and 10° latitude circles of the Northern Hemisphere [16]. Data from observations in January and April 1958--1959 and July and October of 1957-1958 were used for plotting of these curves.

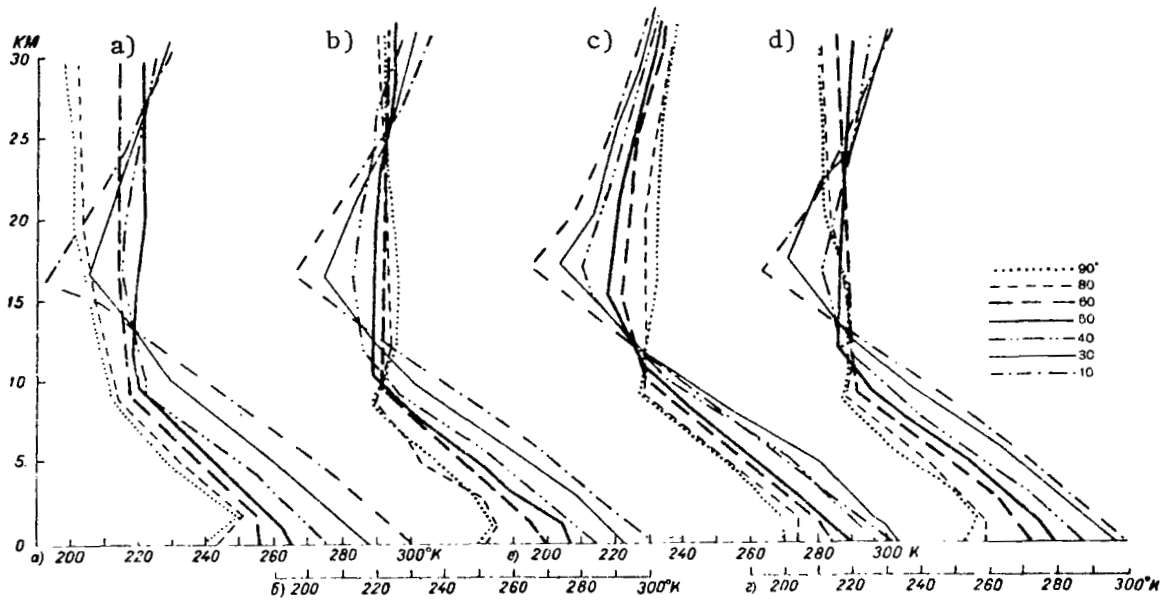


Figure 3.1. Vertical Temperature Distribution over the Pole and 80, 60, 50, 40, 30 and 10° North Latitude Circles (according to V. R. Dubentsov [16]). a, January; b, April; c, July; d, October.

In January (Figure 3.1a) in the high latitudes (80-90° north latitude), due to cooling of the air during the polar night, there is a slight drop in temperature with altitude in the stratosphere up to 30 km. The temperature change in the moderate zone in the stratosphere is characterized by transition from a slight drop with altitude on the 60th latitude to a slight increase to 20 km on the 50th latitude. At the same time the temperature increases rather rapidly with altitude in the lower latitudes above 16-17 km. The opposite vertical temperature distribution occurs in the troposphere. At the lower latitudes the temperature drops more rapidly with altitude than in the middle and polar zones. Here the temperature drop continues at 10 and 30° north latitude to an altitude of 16-17 km, and from the 40th° north latitude to the pole concludes at an altitude of 8-9 km. Because of this vertical temperature distribution the temperature in the winter, in contrast to the other seasons, equalizes in the lower and middle latitudes to the 50th parallel at altitudes of 13 km and about /56
26 km, where all curves intersect at about the same point. During this time of year the temperature in the high latitudes in the entire layer to 30 km is much lower than in the moderate zone. Two specific discontinuities of the temperature curves at 40 and 30° north latitude are noteworthy. The first corresponds to the polar tropopause at an altitude of about 9-10 km, and the second to the tropical tropopause at an altitude of 17 km. Between these levels is located the axis of the subtropical jet stream, which is characterized by a break in the tropopauses [13, 15]. Here the lower polar tropopause is located to the north of the axis of the subtropical jet stream, and the higher tropical, to the south.

In accordance with the described features of the middle vertical temperature distribution in January the meridional component of the horizontal temperature gradient to an altitude of 13-14 km, i.e., to the 150 mb isobaric level, is directed toward the pole from 10° north latitude. The temperature difference here is maximum in the bottom 5 km layer (approximately to the 500 mb level), and it diminishes somewhat at great altitudes. Above the 150 mb level in the latitude zone between 10 and 60° north latitude the horizontal temperature gradient reverses direction--from north to south. The greatest temperature difference here occurs at 16-17 km (near the 100 mb level); at higher altitudes

it diminishes and at 26 km (near the 20 mb level) the horizontal temperature gradient again changes direction. In the high latitudes the horizontal temperature gradient in the stratosphere has the same direction as in the troposphere, and its magnitude nearly quadruples with altitude from the tropopause to 30 km.

The change in direction of the horizontal temperature gradient at the low and middle latitudes at 13-14 km causes the corresponding change in direction of the thermal wind (caused by the horizontal difference in temperatures) from west in the lower layer to east in the higher layer. Consequently there is westerly flow in the troposphere, and it intensifies with altitude, reaching maximum velocity near the 200 mb isobaric level, above which it weakens. In the tropical zone (between 10° and 30° north latitude) there is even east wind above 200-100 mb. In the high latitudes, where the direction of the horizontal gradient remains the same from the earth's surface to 30 km, the velocity of the prevailing west wind increases with altitude by measure of increasing horizontal temperature gradient above the tropopause.

The summer distribution of temperature with altitude (Figure 3.1c) differs substantially from the winter distribution. The highest mean temperatures during the summer solstice, both near the earth surface and in the troposphere, approximately to 12 km, occur considerably to the north of the equator, between 20 and 30° north latitude. The air temperature at 30° north latitude is considerably higher than at 10° north latitude. Due to strong heating of the polar /57
stratosphere at this time of the year the air temperature here does not drop with altitude as in the winter, but rises slightly. Because of these conditions all temperature curves in Figure 3.1c intersect at an altitude of 12-13 km (somewhat above the 200 mb isobaric level), which indicates equalization of the air temperature over the northern hemisphere at the given altitude. The existence of the temperature equalization level at 12-13 km is explained by the varying altitude of the tropopause and unequal vertical temperature gradients in the stratosphere at the high and low latitudes. In the polar region (90-80° north latitude) the temperature above the tropopause rises with altitude an average of 0.4° per km, and in the tropical zone (at 10-30° north latitude) the temperature rise reaches an average of 2.3° per km.

Above 13 km the horizontal temperature gradient assumes a direction opposite that which it had in the troposphere. The change in direction (reversal) of the horizontal temperature gradient causes a change in the direction of the thermal component of the wind or simply thermal wind. Consequently the velocity of the prevailing westerly flow in the troposphere becomes maximum near the 200 mb isobaric level (12 km), and above this level it weakens and changes direction to the east near the 50 mb level (20.5 km). This feature of summer circulation in the stratosphere is described briefly in Chapter 2.

The April and October temperature distributions with altitude (Figures 3.1 b and 3.1 d) have features in common and also display certain spring and autumn features. In particular, the magnitudes of the vertical temperature gradients in the high latitudes are intermediate during these months between their winter and summer values. In the spring the stratosphere near the pole is almost isothermic, and in the autumn there is a slight temperature drop with altitude.

A graphical representation of the seasonal differences in temperature distribution can be seen by the differences in its summer and winter values at various latitudes in the 10-30 km layer (Table 3.1).

TABLE 3.1. SUMMER-WINTER TEMPERATURE DIFFERENCE. ACCORDING TO V. R. DUBENTSOV [16].

Latitude in deg.	Altitude in km		
	10	20	30
80	15	25	35
60	13	11	25
50	13	0	12
30	8	2	4

In the high latitudes the temperature differences are maximum and increase /58 with altitude. To the south they diminish and become slight in the lower latitudes.

Analysis of the above described seasonal features of vertical temperature distribution prompts the following conclusions:

1) In the high and moderate latitudes acceleration of an aircraft to supersonic velocity will take place under more favorable conditions in the autumn and winter because of the low temperature background and small drop with altitude over the tropopause than in the spring, and in particular the summer, when there is inversion and high temperature in the stratosphere;

2) In the low latitudes, where the tropopause is located at 16-17 km or higher and the difference between the winter and summer temperatures above 10 km is slight, thermal conditions will be favorable for transition to supersonic velocity during ascent over the entire year.

The effect of temperature in all stages of flight of a supersonic aircraft are thoroughly examined in Chapter 4.

3.2. Principal Features of Vertical Wind Distribution

1. Dependence of Change in Wind with Altitude on Horizontal Temperature Gradient

Above the boundary layer of the atmosphere (above 500-1,500 m) wind change with altitude depends chiefly on the horizontal temperature gradient. For conditions of geostrophic air motion and coincidence of the directions of the isotherms and isohypses of the absolute topography of isobaric levels, change in wind velocity with altitude can be determined according to the thermal wind formula [15, 30]

$$\frac{\partial u}{\partial p} = \frac{R}{\zeta p} \frac{\partial T}{\partial n}, \tag{3.1}$$

where u is the geostropic wind velocity, p is pressure, R is the gas constant, ζ is the Coriolis parameter and $\partial T/\partial n$ is the horizontal gradient of mean temperature T of the layer. In the troposphere, beginning at the 500 mb isobaric level, the isotherms usually deviate slightly from the direction of the actual wind. Therefore equation (3.1) approximately characterizes the principles of distribution of wind with altitude in the upper troposphere, as well as in the

stratosphere. The value $\partial u/\partial p$ characterizes the vertical wind shear caused by the horizontal temperature gradient.

According to equation (3.1) the wind velocity increases with altitude in proportion to the horizontal temperature gradient. At this level p, where the /59
horizontal temperature gradient becomes equal to zero, $\partial u/\partial p = 0$, and consequently the wind velocity becomes maximum. If the horizontal temperature gradient above this level has the opposite direction the wind velocity drops with altitude and, at some level, may become equal to zero. The higher the horizontal temperature gradient above the level of maximum wind velocity, the more rapidly it weakens. Above the point of zero velocity the wind acquires the opposite direction compared to the direction in the next lower layer and intensifies with altitude.

The velocity maximums are clearly expressed on vertical wind profiles constructed on the basis of rawinsonde data at those points above which are located the jet streams. Examples of vertical wind profiles with sharp velocity maximums in the region of the jet streams are given in Chapter 2. The structure of the jet streams, their geographic distribution and other characteristics are examined in greater detail in special monographs and articles [9, 12, 15, 21, 26, 29-31].

Great ageostrophic deviations of the wind take place in the jet streams. Therefore there may be great deviations of the actual wind distribution with altitude from the distribution determined according to thermal wind formula (3.1). Consequently, the vertical wind displacements below and above the level of maximum velocity in the jet stream, calculated on the basis of observation data, sometimes exceed the thermal displacement two to three times. Such deviations are related to the inertial factor of air movement, which is the result of curvature of flow lines and changes in this curvature with altitude. However, the usual distribution of wind velocity maximums near the level of equalization of temperature of the warm and cold air in the transition layer between the troposphere and stratosphere agrees satisfactorily with the conclusions derived from the thermal wind equation [15, 21, 30].

2. Vertical Distribution of Wind Velocity Maximums in Jet Streams

The vertical wind profile, like the character of change of temperature with altitude, is of tremendous importance in determination of the mode of ascent, cruise and landing approach descent of a supersonic aircraft. In particular, in order to determine the altitude of transition from subsonic to supersonic velocity it is essential to have information concerning the location of maximum wind velocity in the jet stream relative to the tropopause, below and above which are considerable vertical wind shears. Such data are also required for the reverse transition to subsonic speed during landing descent of the aircraft.

Strong winds in the zone of the stratospheric jet stream cause a slight /60
change in maximum range flight time, even if they are head or tail winds at cruise altitudes of supersonic aircraft. Such winds, however, require a considerable change in the number of passengers and the amount of baggage transported by a supersonic aircraft, since the weight of the fuel burned by the engines for flight to maximum range exceeds many times the weight of the payload.

The so-called equivalent wind is used for quantitative evaluation of the effect of wind on the ground speed of aircraft, i.e., on speed relative to the earth's surface, and on over-all flight time. The equivalent wind is calculated by means of a special nomogram or topographic map according to the velocity and direction of the actual wind, and also according to the air speed of the aircraft (its speed relative to air) and ground angle [19-20].

Analysis of the levels and magnitudes of maximum wind velocities exceeding 30 m/sec, filled in on a blank geographical map on the basis of data from radiosonding, makes it possible to determine the spatial location of the axes of the jet streams and their structural features to an altitude of 12 to 14 km. Such maximum wind maps, as we know, are analyzed every day in meteorological centers and are transmitted over facsimile communication channels. They are used extensively for meteorological servicing of the flights of subsonic turbojet and turboprop aircraft. These maps will also be useful for determination of the altitude of the beginning of acceleration of a supersonic aircraft from

subsonic to supersonic velocity. For a cruise of a supersonic aircraft it would be useful to compile and analyze maps of maximum wind in the layer above 14--16 km.

At the present time meteorological centers receive from aerological stations data on wind to altitudes of 20-36 km. The number of rawinsonde observations, however, is still insufficient even for schematic analysis of the stratospheric jet streams.

As an example, a map is given for maximum wind in the 14-38 km layer for 1500 hours on March 14, 1966, over Europe and most of Asia (Figure 3.2). In terms of the amount of data, this may is most complete. It gives an idea of the strong westerly circulation in the stratosphere. The peak velocity of maximum wind, 92 m/sec, is noted in Tura (central Siberian highlands) at an altitude of 26.5 km. Maximum velocity of the west wind, 56-68 m/sec, was recorded in the jet stream over the Caucasus at an altitude of 36 km. This jet stream flows from Asia Minor to the south of Siberia and the Amur Valley.

In the winter the maximum velocities of the wind (MWV) of westerly direction are sometimes much greater in the stratosphere, both over the European and Asiatic USSR. Above Simferopol, for example, at about 9 A.M. Moscow time on /62
January 24, 1963, the MWV at an altitude of 26.2 km was 140 m/sec, and at 3 P.M. on February 29, 1964, over Moscow it was 126 m/sec at 28.6 km. Over Novosibirsk at about 3 P.M. Moscow time on February 29, 1964, the MWV was 197 m/sec at 24.6 km. However, all these maximum wind velocities were located considerably higher than the cruising altitude of the first supersonic transports. Consequently, they will have no effect on their flight.

In the summer it is possible to approximate the position of the maximum wind level in the eastern stratospheric jet stream in the tropics over the southern periphery of Asia and the Philippine Archipelago. In this part of the Northern Hemisphere the stated jet stream apparently is stronger. The level of maximum wind in its vicinity is found at altitudes of 20-35 km. Beneath it, at altitudes of 14-16 km, tropospheric maximum velocity of the east wind is often encountered.

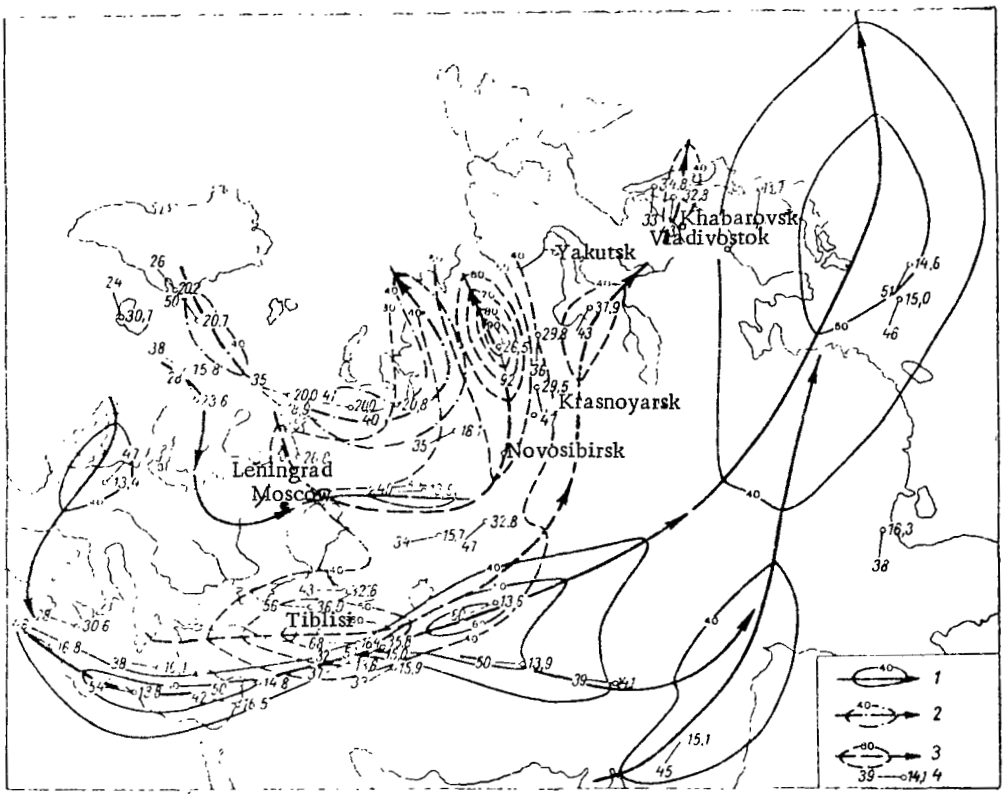


Figure 3.2. Jet Streams on Map of Maximum Wind Above 150 mb. 1500 hours, March 14, 1966. 1, Axis of jet stream and 40 m/sec isotach at altitudes of 14-16 km; 2, Same at altitudes of 20-24 km; 3, Axis of jet stream and 60 m/sec isotach at altitudes of 26-38 km; 4, Direction of maximum wind (arrow), its velocity (m/sec) and altitude (km).

Curves of distribution of maximum wind velocity levels, plotted according to radiosonde data for 1963-1964 at Moscow, Khabarovsk and Novosibirsk [24], give an idea of the altitudes of jet stream axes in the upper troposphere and stratosphere in the Moscow-Khabarovsk lane. Maximum wind velocities over Moscow and Novosibirsk (Figure 3.3) are most often encountered in the layer between the 400 and 200 mb levels (7-12 km) and over Khabarovsk (Figure 3.4) between the 400 and 150 mb levels (7-13.5 km). In the overlying layer, between 150 and 50 mb (13.5-20.5 km) maximum wind velocities are rare. When these maximum wind velocities exceed 50 m/sec (180 km/hr) they are encountered most

frequently above the first two points in the layer between 500 and 200 mb, and between 500 and 100 mb above the third point (Table 3.2). The number of cases with such high maximum velocities increases from west to east. The over-all recurrence of maximum wind velocities of 31 m/sec and greater decreases notably over Moscow and Novosibirsk in the 99-30 mb layer (16-24 km) and over Khabarovsk in the 49-30 mb layer (21-24 km), and increases somewhat again in the 29-10 mb layer (24-31 km). The reduction in a number of wind velocity maxima over the specified points in the 16-24 km layer agrees with the position of the maximum velocity of the westerly winds in the moderate latitudes during winter at an altitude of about 20 km (according to data from the meteorological rocket network) [35]. This minimum is thoroughly visible on the background of constant wind velocity maxima of the same direction near the tropopause and in the mesosphere at an altitude of about 60 km [16, 41]. In the summer the maxima of the west wind are concentrated principally in the upper troposphere, and in the mesosphere rocket sounding reveals maximum velocity of the east wind near the tropopause, at an altitude of about 50-60 km [16, 41, 44].

Above the jet streams, the axes of which are located in the upper troposphere, there are considerable vertical wind shears in the lower stratosphere. The mean vertical wind shears above the jet stream axis, like those below it, depend on the maximum wind velocity (Figure 3.5). As maximum /63
velocity increases the vertical wind shear increases. Here it is somewhat greater above the level of maximum velocity than below it [23]. In individual cases the vertical wind shear can exceed the main values in Figure 3.5. The graph of the correlation relation of vertical wind shears and maximum velocity is also plotted in [10]. Usually the vertical wind dis-
shear is maximum in approximately a 1 km layer below and above the maximum velocity level and diminishes below and above it approximately exponentially by measure of distance from it. Nevertheless the vertical shear is still /64
great in a layer of up to 2 km thickness above the maximum wind velocity level.

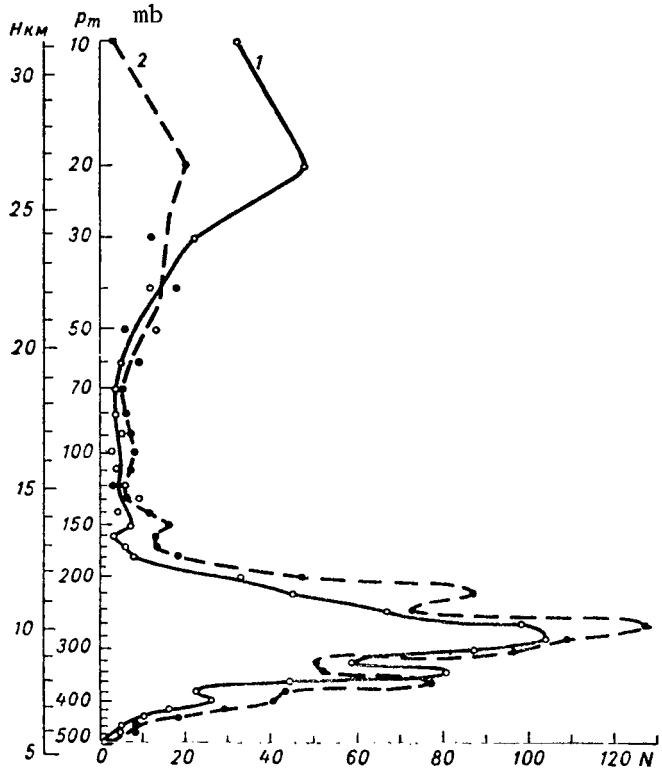


Figure 3.3. Curves of Distribution of Maximum Wind Velocity Levels in the Upper Troposphere and in the Stratosphere over Moscow (1) and Novosibirsk (2). N, Number of cases.

Over southern USSR, where subtropical jet streams prevail in the summer, the distribution of wind velocity maxima with altitude is somewhat different than in the middle zone. Here, according to the data of 15 stations for the summer months of 1957--1959, their repetition is maximum at 12 km. Below 9 km and above 14 km it decreases sharply [14]. Above eastern Pamir the maximum wind velocities during the same months of 1957-1958 occurred primarily in the layer between 9 and 16 km [32]. In the middle vertical cross-sections along the 40, 60 and 75° East longitude meridians for July and October of 1958-1959, constructed according to daily maps of maximum wind, the maximum velocities were concentrated in the 9-14 km layer between 42.5 and 32.5° north latitude [31]. In January and April the greatest number of velocity maxima occurred in the same latitude zone in the 8-12 km layer, i.e., 1-2 km lower than in summer and autumn. Here, the region of high repetition of the axes of the jet streams of the moderate latitudes is located between 8 and 10 km in January, and above them, between 10 and 12 km, is located a region of /66 high repetition of the axes of subtropical jet streams. The number of the latter is smaller in this latitude zone during the winter due to their frequent displacement to the south under the influence of sharp cold invasions from European USSR and from Siberia.

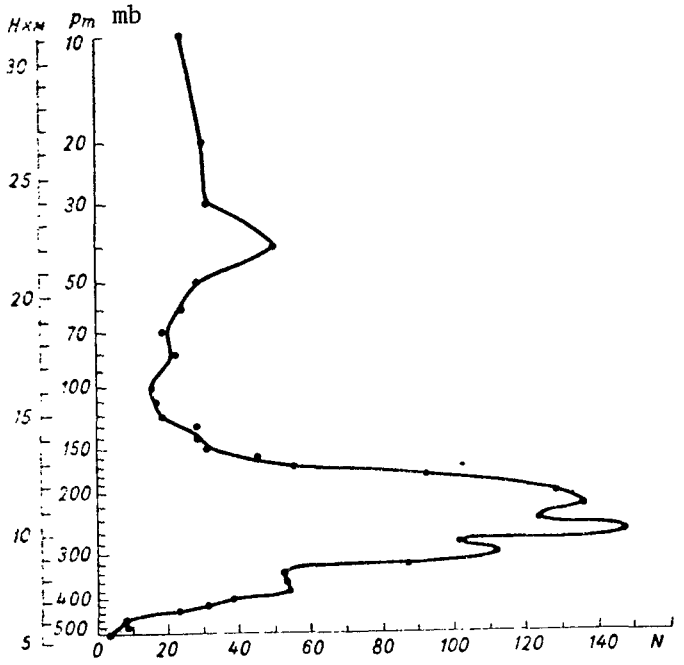


Figure 3.4. Curve of Distribution of Maximum Wind Velocity Levels in Upper Troposphere and in Stratosphere Over Khabarovsk. N, Number of cases.

Because of the above-described features of vertical distribution of wind velocity maxima in January the zone of mean wind velocities exceeding 30 m/sec at the 100 mb isobaric level encompasses the subtropics and tropics of the northern hemisphere, and at the 30, 20 and 10 mb levels, the middle latitudes [16]. At the same time the mean wind velocities near 30 m/sec occur only in the middle latitude zone at the 50 mb level. In the direction toward the pole and equator the mean velocities decrease to 10 m/sec. In July mean velocities less than 30 m/sec at the 100, 50 and 30 mb levels encompass nearly the entire hemisphere, with the exception of the tropical zone. Here the mean velocities in the polar region are less than 10 m/sec.

Thus it can be concluded that supersonic passenger planes will cruise at altitudes of 18-20 km in the layer with maximum wind velocities. Correspondingly the equivalent wind that influences flight will be less than in the layers lying below and above. Only in the winter during flight in the middle zone from east to west will the velocity of the equivalent wind exceed 30 m/sec in some segments of the trip.

It will be necessary, for determination of the point of the level of transition to supersonic velocity, not only to select in the ascent zones the layer with the lowest air temperature, but also to consider the location of the

level of maximum wind velocity and magnitudes of the vertical wind shear in a layer about 2 km thick above this level. To avoid sonic boom, the conditions of which are described in Chapter 4, transition to supersonic velocity in the ascent zone must be made in the layer with the lowest temperature near the tropopause and to 1-2 km above the level of maximum wind velocity, where there will be no considerable vertical wind shears.

TABLE 3.2. REPETITION (NUMBER OF CASES) OF MAXIMUM WIND VELOCITIES IN JET STREAMS AT VARIOUS LEVELS OVER MOSCOW, NOVOSIBIRSK AND KHABAROVSK [24].

Level in mb	Maximum wind velocity, m/sec						
	31 - 40	41 - 50	51 - 60	61 - 70	71 - 80	> 80	> 31
				Moscow			
500-200	384	190	43	10	3	1	631
199-100	36	14	3	1	-	-	54
99-50	11	8	2	1	-	-	22
49-30	16	10	5	2	-	-	33
29-10	22	21	14	6	4	2	69
9-4	4	3	2	2	1	-	12
500-4	473	246	69	22	8	3	821
				Novosibirsk			
500-200	369	240	81	44	7	10	751
199-100	41	24	8	7	4	-	84
99-50	4	14	3	1	1	3	26
49-30	5	3	5	-	4	2	19
29-10	7	5	10	5	-	3	30
500-10	426	286	107	57	16	18	910
				Khabarovsk			
500-200	438	219	128	48	23	22	878
199-100	183	94	34	16	2	4	333
99-50	59	13	5	-	-	-	77
49-30	15	14	2	-	-	-	31
29-10	27	23	3	1	1	1	56
9-3	5	3	1	-	1	1	11
500-3	727	366	173	65	27	28	1,386

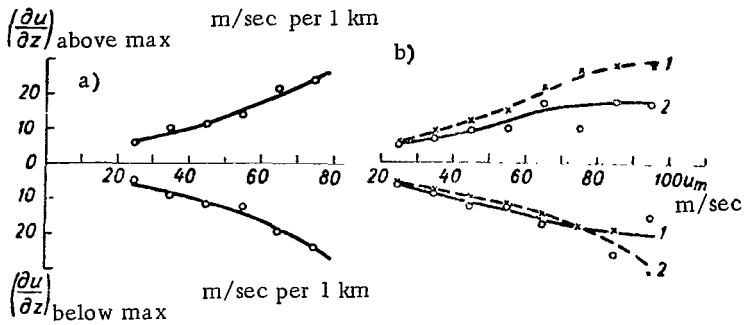


Figure 3.5. Average Vertical Wind Shears in 1 km Layer Below $\partial u/\partial z_{\text{below max}}$ and Above $\partial u/\partial z_{\text{above max}}$ Maximum Velocity Level (a) Over Moscow and Novosibirsk (b) Below 300 mb (1) and Above 300 mb (2).

Since January 1, 1968, additional information concerning vertical wind shear in the layers 1 km below and 1 km above the level of the maximum wind exceeding 30 m/sec has been transmitted in the USSR from sounding points by the new KN-03 and KN-04 codes [17, 18].

Not only current information, but also precalculation of the level and the velocity of maximum wind in the jet streams, as well as of the vertical wind shears below and above the jet stream axis, are important in determination of the conditions of ascent and the level of transition to supersonic velocity.

A method was proposed recently for predetermination of the altitude (or pressure) and velocity of maximum wind according to formulas, into which are substituted the geopotential altitudes and velocities of the wind from forecast /67 maps of the topography of two isobaric surfaces, between which lies the jet stream axis [22]. Precalculation is also possible in the case when the jet stream axis is located above both isobaric surfaces for which forecast maps are constructed.

Using the altitudes and maximum wind velocities precalculated at the centers of the regular network and filled in on a clean map blank it is possible to outline the isotachs and determine the future location and altitude of the jet stream axis.

The maximum wind forecast maps thus obtained are used as the basis of the procedure of forecasting vertical wind shears in the jet stream, of which pilots must be informed [23]. The data required for this purpose are determined

with the aid of the relationship between maximum wind velocity and the arithmetic means of the absolute values of vertical wind shears in the 1 km layer below and above the maximum velocity level.

The above-described method makes it possible to precalculate the level and velocity of the maximum wind in the jet streams, the axis of which is located not higher than 12-14 km, since forecast maps of the topography of isobaric surfaces are presently constructed only to the 200 mb level.

In future years it will be possible, by measure of increasing regularity of radio wind observations up to the rawinsonde level, to study the jet streams in the 16-30 km layer to the same extent that they had been investigated over the last 10-15 years in the 5-15 km layer. This will permit precalculation of the velocity and vertical displacements of the wind in the zone of stratospheric jet streams.

3. Layer of Wind Reversal in the Stratsophere During Spring-Summer Period

It was pointed out in Chapter 2 that in the summer anticyclonic circulation occurs in the stratosphere over much of the northern hemisphere. Consequently, east wind prevails above 20 km from April-May through September over almost the entire hemisphere. Thus the wind reversal occurs during the specified period, i.e., westerly flow of the troposphere and lower stratosphere is replaced by easterly flow in the remaining thickness of the stratosphere. In the transition layer the velocity of the prevailing west wind decreases with altitude with the result that in the summer (see §3.2.1 on p. 61) the horizontal temperature gradient below the tropopause extends from the equator to the pole, and above it, in the opposite direction, from the pole to the equator. Theoretically the wind velocity at the reversal level should be equal to zero. Actually, near the reversal level there are slight, directionally unstable breezes and it is rarely calm [1-6, 8, 25]. Over Moscow, Novosibirsk and /68
Khabarovsk, for example, the minimum wind velocity in the layer of transition from west to east does not exceed 5 to 6 m/sec in most cases, but sometimes the wind weakens to a calm [24].

The transition layer between the westerly and easterly breezes or calm in the stratosphere is called the vetropause [1-3] or velopause [4-8]. The thickness of the transition layer varies from 1 to 5-7 km and averages 2-3 km. In the transition layer the wind rotates clockwise or counterclockwise with altitude. When rotation is clockwise winds of the northern quarter predominate, and when counterclockwise, winds of the southern quarter. The upper boundary of the above-stated transition layer, i.e., the lower boundary of stable stratospheric wind of the eastern quarter (45-135°) is used in [1-8] as the height of the vetropause or velopause. In jet aviation meteorological service the altitude of the lower boundary of the stable east wind in the stratosphere has come to be called the level of wind reversal. In the ensuing discussion, therefore, the velopause will be identified with this level.

In the middle latitudes the wind reversal level is usually located between the 70 and 20 mb isobaric levels, i.e., at altitudes of approximate 18 to 26 km [1-3, 5-8, 21, 22, 36]. It descends toward the pole [38] and toward the /69
tropics [37]. Data concerning altitudes of the wind reversal level are also found in [16, 33-39]. The curves of distribution of the wind reversal level with altitude in the moderate latitudes--in Moscow, Novosibirsk and Khabarovsk [24], are illustrated in Figure 3.6. The maximum of cases for these curves falls at the 40 mb level, or at 22 km. At the wind reversal level the wind velocity over these points is somewhat greater than the minimal velocity in the transition layer from westerly to easterly flow and usually varies from 1 to 8 m/sec, although in some cases it reaches 9-10 m/sec and beyond.

The wind reversal level sometimes coincides with the minimum velocity level (in 16% of the cases over Moscow and in 13% of the cases over Novosibirsk), but often it is higher (in 67% of the cases over Moscow and 71% of the cases over Novosibirsk). Only in a small number of cases (16-17% over the above-mentioned points) is the wind reversal level lower than the minimum velocity level [24].

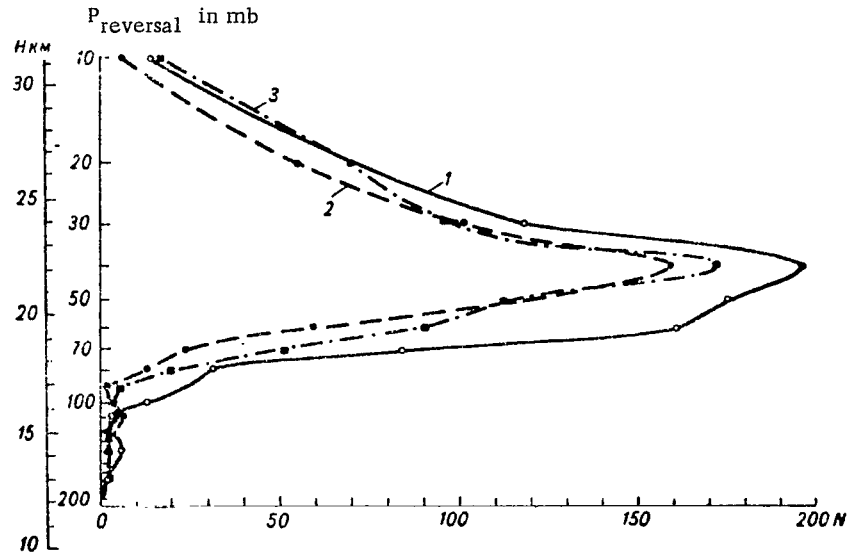


Figure 3.6. Curves of Distribution of Cases of Wind Reversal Level in the Stratosphere at Various Isobaric Levels (Altitudes) Over Moscow (1), Novosibirsk (2) and Khabarovsk (3).

The lower boundary of the easterly flow in the stratosphere, or reversal level, generally drops from April-May to June-July and rises from July to September [2, 5, 7, 8, 24]. The minimum altitude of the reversal level occurs near the summer solstice (June 21-22) or somewhat later. Thus, June and July are the months of the lowest location of the lower boundary of the eastern flow in the stratosphere, according to average data. Thus, for instance, the mean monthly altitude of the level of reversal of the wind to the east over Khabarovsk [24] drops 5 km between April and June (Figure 3.7); it is about the same in June as in July, but then it increases 7 km by September. Over Moscow and Novosibirsk, however, the mean monthly altitude of the wind reversal level was less in April than in May. Under this condition there was a substantial drop of the wind reversal level from May to June. The lower position of the wind reversal level over Moscow at the beginning of spring was noted earlier during the period of the International Geophysical Year [25]. The mean monthly altitude of the wind reversal level, according to data from a number of

stations (Figure 3.7), over Kazakhstan and Central Asia drops from May to June or July, and then rises most substantially from July to September [2].

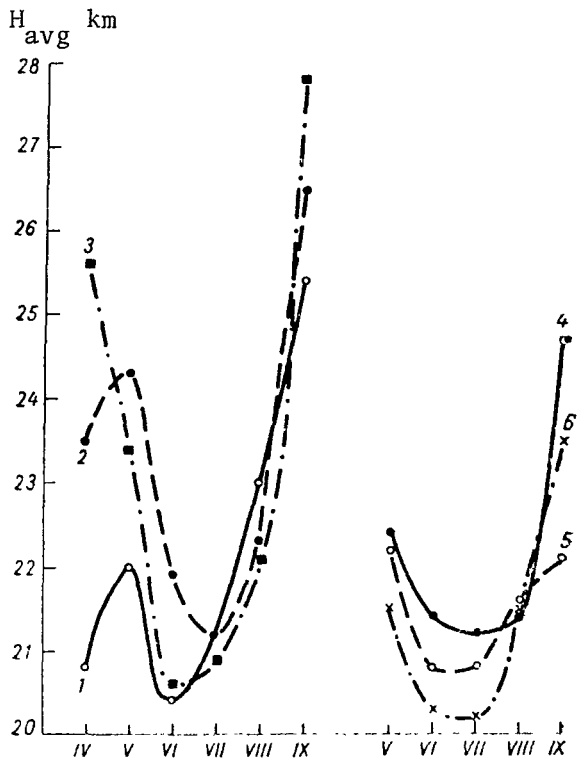


Figure 3.7. Curves of Altitude of Wind Reversal Level in the Stratosphere During April-September over Moscow (1), Novosibirsk (2), Khabarovsk (3) and During May-September over Uralsk, Aktyubinsk, Kustanay (4), Tashkent, Dzhalal-Abad, Dushanbe (5), Krasnovodsk, Gasan-Kuli, Ashkhabad (6). (1-3 according to N. V. Petrenko [24]; 4-6 according to R. Bilyalov [2]).

According to the above-mentioned average data, the minimal altitude of the wind reversal level in June or July in Figure 3.7, both in the middle zone and over southern USSR, falls between 20 and 21 km. In May the mean altitudes of the wind reversal level in the middle zone are 22-24 km, and 1 km lower to the south; 25-28 km in the middle zone in September and 23-24 km over southern USSR.

The level of wind reversal to the east in the spring over the southern Hindustan Peninsula, in the region of Hyderabad, is 19 km [37]. The maximum easterly velocity here occurs at an altitude of about 26 km, and at 27-32 km (near the ceiling of the radio sondes) westerly flow again occurs in certain cases.

For determination of flight /70 conditions at altitudes of 18-22 km on various long-range airlanes is is useful to have an idea of the spatial position of the lower boundary of the summer stratospheric east wind, i.e., wind reversal level.

On the mean monthly topographical maps of the lower boundary of the east wind (velopause) over the northern hemisphere [7, 8] the isohypses do not have strictly zonal distribution (Figure 3.8), but their configuration changes slightly from May to August. During any of the four months (May, June, July and August) the lower boundary of the east wind over the polar region forms a funnel. The wind reversal level rises gradually in the direction of the middle latitudes. Along the middle latitude zone there is a belt in which the lower boundary of the east wind is in its highest position. It is located to the north of axis of the planetary subtropical jet stream. /71

In May (Figure 3.8a) the lowest altitude of the wind reversal level near the pole is 16.5 km. In the maximum altitude zone of the wind reversal level, along the 40-45° north latitude zone, there are four domes with altitudes of 22.5 to 21.8 km. South of the zone its highest position again drops, and in a 15-20° North latitude zone its altitude is about 19 km.

In June (Figure 3.8b) the basic features of the position of the wind reversal level remain the same as in May. However its mean altitudes are about 1 km lower everywhere. In the polar region there is a 16.0 km isohypse; in the middle latitudes are domes with reversal level altitudes of 21.5-21.0 km.

In July (Figure 3.8c) the altitudes of the wind reversal level over the entire hemisphere are even lower. Over the Arctic is a 15.0 km isohypse; the isohypses in the middle latitudes are 20.5-21.0 km and 16.0 km in the tropics. Consequently, during the period of greatest development of the polar anticyclone in the stratosphere the altitudes of the reversal level reach minimal values. Its maximum altitude domes in the middle zone are displaced toward higher latitudes, and accordingly to a more northerly position of the subtropical jet stream during the second half of the year.

In August (Figure 3.8d) the altitudes of the wind reversal level are higher. Thus, the altitudes increase to 22.0-22.5 km in the zone of its highest position. The axis of this zone is now located somewhat farther north than during the preceding months (along the 50-52° North latitude). Its more northerly position agrees well with the shifting of the axis of the subtropical jet stream to the north. Some increase in the altitude of the wind reversal level also occurs in the polar region (16.0-16.5 km).

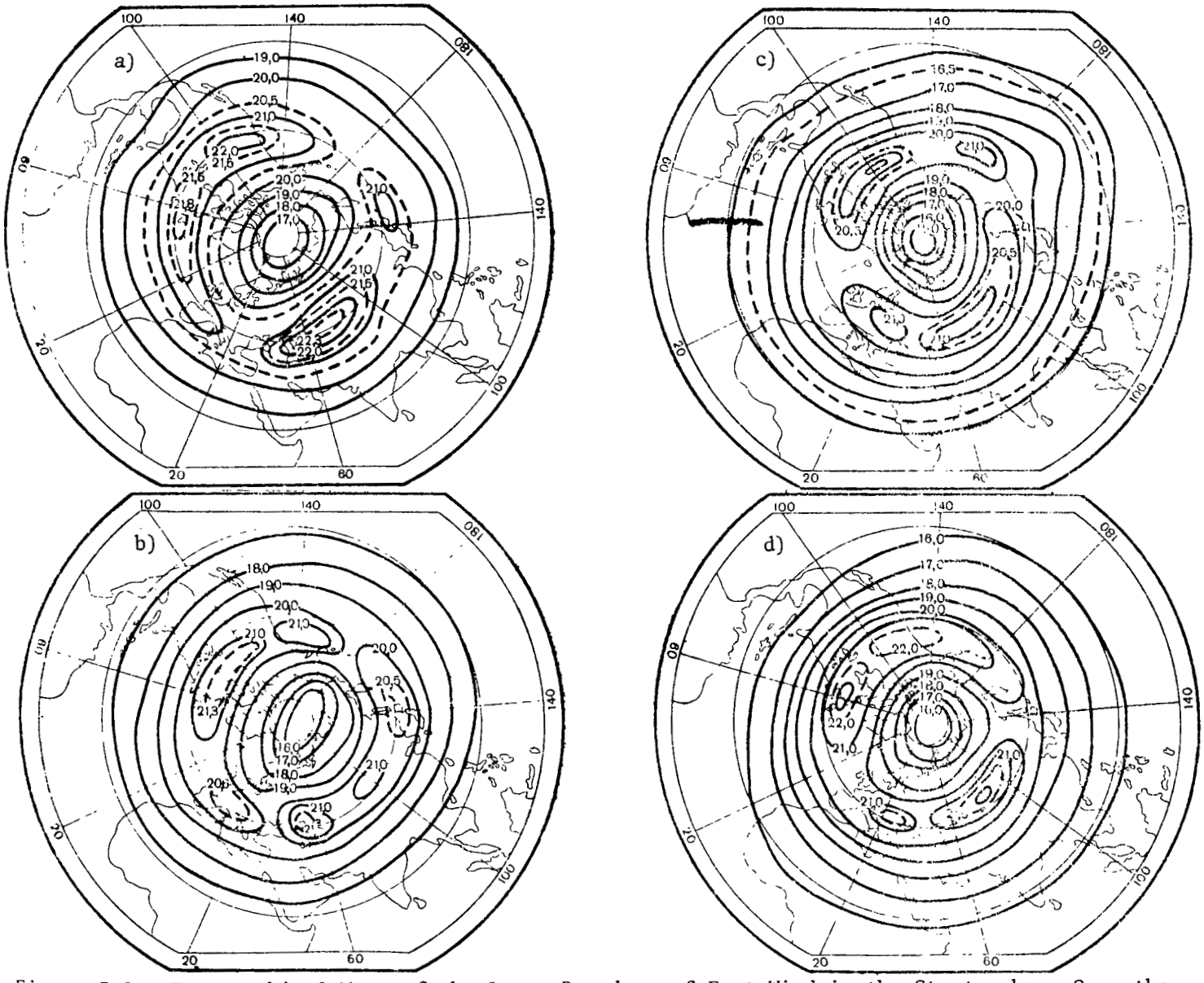


Figure 3.8. Topographical Maps of the Lower Boundary of East Wind in the Stratosphere Over the Northern Hemisphere in 1959 (According to I. V. Bugayeva [7]). a, May; b, June; c, July; d, August.

In meridional cross-section of the mean zonal circulation in July constructed for the Northern Hemisphere regardless of the longitude of the location, the western component of the wind above 12 km diminishes to zero in the middle latitudes and reverses to the eastern component at 19 km [11].

It will be possible in future years, obviously, to use data concerning the summer east wind in the stratosphere of the Southern Hemisphere to insure supersonic flight. Above Antarctica, according to rawinsonde observations at Mirnyy (on the Antarctic Circle) the date of the beginning of easterly wind in the summer stratosphere varies approximately between the first of November and the first half of December, and at Vostok Station (on the 78°27' south latitude /74
according to insufficiently regular rawinsonde observations to high altitudes) approximately between mid-December and the first day of January. The date of completion of the summer east stratospheric wind at these points varies from the beginning to end of February [4]. The average location of the wind reversal level during individual years and during the middle months of summer--December and January--varies over Mirnyy from 18.5 to 24 km. The average altitude for six summers is 20.5 km. This is approximately 2 km higher than at the Arctic Circle. The average altitude of the wind reversal level at Vostok Station for five summers, and during individual months, is quite stable and amounts to about 16.0-16.5 km. This corresponds approximately to its altitude above the analogous latitudes in the Arctic. Hence, it follows that the funnel formed by the lower boundary of the layer of the east stratospheric wind is more distinct during the summer over Antarctica than over the Arctic.

The time of change of stratospheric circulation from western in winter to eastern in summer is determined, in general, by the influx of solar radiation, but lags behind it. In the middle latitudes of the Northern Hemisphere, according to data from the rocket meteorological network, the east wind in the upper stratosphere is established approximately 30 to 40 days after the vernal equinox, i.e., between April 1 and May 1. The maximum strength of summer eastern circulation is reached in the middle of July, approximately 26 days after the summer solstice. The transition, however, to winter westerly circulation takes place almost immediately after the autumnal equinox, about the 22nd

of September [43, 44]. The polar summer seasons in both hemispheres, according to the same data, are shorter and apparently begin earlier.

Easterly circulation of the summer hemisphere extends out to at least the mesopause (approximately 80 km) and penetrates into the winter hemisphere. Here the maximum east winds at a level of approximately 50 km can be discovered in the lower latitudes of the equatorial zone of the summer hemisphere. The invasion of east wind of the summer hemisphere into the lower latitudes of the winter hemisphere results in the development of a zone of strong wind shear between these opposing circulation systems. The maximum meridional wind shear, according to average data, occur in the region of latitudes from 10 to 20°. Here the meridional wind velocity gradient reaches several meters per second per degree latitude [44].

Spring transition from westerly to easterly wind in the stratosphere caused by warming in the polar zone and in the middle latitudes, takes place at an altitude of approximately 35 km [40, 44]. Then the easterly flow spreads downward at a rate of 1-5 km a day. In the autumn the reverse transition from easterly to westerly wind begins at an altitude of 50-60 km due to cooling of the mesosphere and the stratosphere from the pole toward the middle latitudes. Then the westerly wind descends more uniformly than in the spring, at a rate of /75 approximately 1 to 4 km per day. On about the 15th of September (on an average), the east wind is completely replaced by west wind through the entire thickness of the stratosphere.[40, 44].

According to radiosonde observations to altitudes of 25-35 km, the time of appearance and disappearance of east wind in the middle stratosphere varies from year to year. Over central Asia, for instance, east winds occurred in the stratosphere in 1958 beginning with the second half of March, and in 1959 starting in April [2]. As mentioned in Chapter 2, these fluctuations are attributed to the earlier or later formation and destruction of the summer polar stratospheric anticyclone at isobaric levels from 10 to 50 mb under the influence of the features of large-scale disturbances in the troposphere and stratosphere and possibly under the influence of other factors as yet unanalyzed. Thus, for instance, in 1958 the formation of the summer stratospheric

anticyclone over the polar basin of the northern hemisphere occurred on about 10 May and in 1959, between the 10th and 15th of April [27].

Apparently the time of appearance and disappearance of the east wind in the stratosphere depends not only on fluctuations in the influx of solar energy and other cosmic events, but also on features of the preceding over-all circulation of the atmosphere.

The dates of the first and last recordings of east wind in the stratosphere, according to radiosounding data, differ notably not only from year to year, but also in the various physico-geographic regions during the same year. This is related to the asymmetry of the position of the stratospheric anticyclone relative to the latitude circles, particularly at the beginning of its formation in the spring and during the period of destruction in the autumn. According to the stated reason, the east wind was first recorded at Khabarovsk in 1963 at an altitude of 29 km (13 mb) on April 1, and in Moscow at an altitude of 30 km (12 mb) on about April 17 [24]. At Novosibirsk, located between Moscow and Khabarovsk, the later date of the first recording of east wind (April 21) can be explained by the low altitude of rawinsonde observations during this period. Here the east wind was recorded four days later than at Moscow, after its descent to an altitude of 23 km (34 mb). In 1964 the east wind was first recorded at Khabarovsk on March 5th and in Moscow on March 22nd, i.e., 25 days earlier than in the previous year at the same altitudes (29-30 km, i.e., approximately 12 mb). At Novosibirsk in 1964 the later date of the first observation of east wind (April 1) can be attributed to the same causes as in 1963. It is noteworthy that the last dates of recording of east wind in 1963, ignoring the Novosibirsk data (5 September), differed by only seven days: September 22 in Khabarovsk and September 29 in Moscow. In 1964, the same dates differed by 3-7 days: September 13 in Novosibirsk, September 17 in Khabarovsk and September 20 in Moscow. Thus, it can be assumed that during the two examined years the east wind was noted the last time from the 13th to the 29th of September. Here the altitude of the east wind changed from 22 to 29 km (45-14 mb). Comparatively slight deviations of the time of autumn replacement of circulation in the stratosphere from the average date were noted earlier during other years as /76

well. The dates of spring replacement of circulation, on the other hand, varied more compared to the dates of the autumn replacement of circulation [16].

Since supersonic transports will have an average flight altitude of approximately 20 km, the USSR plans to obtain current information concerning the level of wind reversal in the stratosphere from west to east. This information will be included into additional groups of codes for transmission of temperature--wind and wind sounding data [17, 18]. In this group will be specified the pressure or altitude of the eastern quarter at the lower boundary of the stable wind layer (40-140°), and also the direction and velocity of the wind at the lower boundary of this layer. The information concerning the level of wind reversal entered on the AT_{50} map will enable a meteorologist to determine in what segments of the air lane this level will be higher or lower than the 50 mb isobaric surface during the given period of observations.

During analysis of the variations in the reversal level according to radio sounding data the substantial error of determination of wind direction at velocities of 5-6 m/sec and less must be kept in mind. In addition, above 16 km (100 mb) the average wind is measured in layers 1.5 to 2.0 km thick [25]. Accordingly the altitude of the reversal level can be established with the same error. Therefore the correct representation of the character of active changes in the reversal level can be obtained when their magnitude, according to observation data, exceeds 2 km. The stated circumstance complicates determination of dependences for forecasting the level of wind reversal in the stratosphere. Only the first steps have been taken in this direction.

A linear dependence of the reversal level altitude on the altitude of maximum wind velocity occurring above 12 km has been found for the southern part of the USSR [2]. It is represented as the empirical equation

$$H_n = H_{n-1} + 0.1(1 + n),$$

where $n = H_m - 11 \geqslant 1$; $H_{n-1} = 20.5$ km at the altitude of maximum wind velocity $H_m = 11$ km.

The dependence of the reversal level altitude on wind velocity along the jet stream axis, not exceeding 50 m/sec, has also been established. The stronger the subtropical jet stream to the stated velocity limit, the higher the reversal level. At velocities exceeding 50 m/sec, however, elevation of the reversal level as a function of the altitude of the jet stream axis takes place more slowly.

If the subtropical jet stream axis is located at an altitude of 13 km the reversal level will be located at the same altitude, 21 km, regardless of the wind velocities in it.

It should be pointed out in conclusion that wind velocity during summer on the 50 mb isobaric level, located near the cruise level of the present /77
supersonic transports, usually differs only slightly from the wind velocity at the reversal level. Over Moscow, Novosibirsk and Khabarovsk, for example, the wind velocity at the 50 mb level in the summer varies in most cases from 1 to 10 m/sec [24]. The mean wind velocity at this level in July over the polar and moderate zones of the Northern Hemisphere is less than 10 m/sec [16]. Only in the tropical zone is it 20-30 m/sec. Thus, supersonic transport flight in the summer will take place, as a rule, in the zone of weak winds of easterly or unstable direction. Changes in these winds from the initial period to the next period of sounding are generally small. However, they may be considerable when large cyclones and powerful anticyclones, reaching up to the 100 mb level and beyond, form in the troposphere. Toward the end of summer these changes in the east wind are also determined by evolution of the fading stratospheric anticyclone.

CHAPTER 4

INFLUENCE OF TEMPERATURE AND WIND ON SUPERSONIC TRANSPORT FLIGHT

/78

4.1. Features of Supersonic Flight

As the speed of aircraft increases to supersonic the effect on it of the temperature of the ambient air increases and the significance of the wind decreases.

Air temperature has a tremendous effect on the thrust of a jet engine, and consequently on the rate of ascent of an aircraft, its acceleration to supersonic velocity and on the speed and altitude of cruise flight.

The thrust of a jet engine increases as the temperature of the ambient air drops and decreases as the temperature rises [2, 17]. In the real atmosphere, therefore, which is considerably warmer than the standard atmosphere, more time is required to reach the altitude where supersonic acceleration begins due to weakening of the thrust of the engines, and acceleration will be less. Consequently much more fuel will be consumed by the time the aircraft reaches cruise altitude than in the case of the distribution of temperature with altitude in the standard atmosphere (SA). A very high temperature in the stratosphere, compared to its value in the SA (-56.5°C), in conjunction with kinematic heating of the leading parts of the aircraft, is the cause of limitation of supersonic velocity at cruise altitude. The forced reduction of cruise velocity causes a reduction of flight altitude, which leads to additional fuel consumption.

The maximum permissible (limiting) velocity and flight altitude are established for each type of aircraft depending on its performance characteristics and the air temperature.

The Mach number is an important criterion for supersonic flight. It gives an idea of the influence of the compressibility of air on the motion of an aircraft and is found from the relation

$$M = V/a, \qquad (4.1)$$

where V is the airspeed of the airplane, a is the speed of sound (velocity of propagation of the sound waves excited by the aircraft). /79

The process of deformation of an airstream during the propagation of sound waves in it is usually regarded as adiabatic [18]. The speed of sound, therefore, depends only on the air temperature and it is calculated using the equation

$$a = \sqrt{\chi RT}, \tag{4.2}$$

where χ is the adiabatic index (ratio of specific heat capacity of pure air at constant pressure and a 0°C to the specific heat capacity of the same air in constant volume), R is the specific gas constant of dry air, T is the absolute temperature (in degrees Kelvin).

By substituting into the equation $\chi = 1.402$ and $R = 287.04$ $m^2/sec \cdot degree$ [14], we obtain

$$a = 20.1\sqrt{T}\ m/sec. \tag{4.2a}$$

At an air temperature of 0°C the speed of sound is a = 332 m/sec, or approximately 1,200 km/hour.

From relations (4.1) and (4.2a) we obtain:

$$V = 20.1\ M\sqrt{T}\ m/sec, \tag{4.3}$$

or

$$V = 72.4\ M\sqrt{T}\ km/hour. \tag{4.3a}$$

In the case $M < 1$ motion is called subsonic, and when $M > 1$ it is supersonic. The compressibility of the air flowing past an aircraft is greater during supersonic flight than during subsonic flight. As the Mach number increases, i.e., as the ratio of airspeed of the aircraft to the speed of sound increases, the frontal resistance of the air increases.

The following phases of the flight of supersonic aircraft have come to be distinguished in aviation [23]: take-off and initial ascent phase (gaining altitude) from subsonic velocity corresponding to Mach numbers up to 0.8; transsonic phase--period of continued ascent (gaining altitude) with increasing velocity from subsonic to supersonic in the Mach number range of 0.8 to 1.2; cruise phase--period of flight with gradual ascent and with velocity increasing to the limiting value at cruise altitude, which corresponds to Mach numbers from 1.2 to the maximum tolerable (limiting) velocity for each type of aircraft; deceleration and landing phase--concluding a stage of flight, starting with reduction of velocity from supersonic to subsonic.

Sometimes the near sonic phase is limited to Mach numbers of 0.9-0.95 to 1.5-1.6.

A temperature exceeding standard in the subsonic, and particularly in the near sonic phases, causes the greatest increase in fuel consumption. In the cruise phase a temperature exceeding the standard temperature limits the maximum flight speed. /80

4.2. Velocity and Altitude as Functions of Temperature

With equation (4.3) it is possible to calculate the maximum airspeed of an aircraft according to the maximum Mach number established for it at the cruise altitude and at the temperature of the air flowing past the aircraft. Here, however, it must be borne in mind that due to deceleration of the airstream at the leading parts of the aircraft kinetic energy is converted into heat energy. The amount of kinetic heating of completely stagnated air in the boundary layer of the leading parts of the aircraft is determined according to the familiar equation [2, 16].

$$\Delta T = V^2/2{,}000 \ , \qquad (4.4)$$

where V is the airspeed of the aircraft in meters/second.

The skin of the aircraft will experience the corresponding heating. The nose of the aircraft, where the kinetic energy of the decelerated airstream is

almost entirely converted to thermal energy, will experience the greatest heating. The leading edges of sweptback and delta wings are heated less due to the breaking down of the airspeed vector into normal and tangential components to the leading edge of the wing. The flight surfaces of the fuselage are heated even less.

Solving jointly (4.3) and (4.4), we obtain

$$\Delta T = 0.2\ M^2 T.$$

Using this relation we find the absolute temperature of kinetic heating of completely stagnated air near the nose of the aircraft[2]

$$T_K = T + \Delta T = T(1 + 0.2\ M^2). \qquad (4.5)$$

It follows from equation (4.5) that when $M = 1$, $T_K = 1.2T$, when $M = 2.2$, $T_K = 2T$ and when $M = 3$, $T_K = 3T$.

The above relations between the temperature of the undisturbed air flow T at the flight altitude and temperature of kinetic heating T_k at all three Mach numbers yield an important practical conclusion: at an airspeed corresponding to M = 2.2 in the stratosphere of the SA, i.e., at an air temperature equal to -56.5°C, the kinetic heating temperature approaches the limit at which the tensile strength of ordinary aluminum alloy is notably reduced [23]. Therefore, /81

[2]Equations (4.4) and (4.5) are extremely approximate for determination of the heating of the aircraft's skin. The air temperature in the boundary layer is known in aerodynamics as the regeneration temperature. The temperature of the skin and other parts of the aircraft's structure can differ considerably from the regeneration temperature. Due to convective heat transfer some of the heat from the boundary layer is transmitted to the skin, and from the skin to the within the structure. Part of the heat is radiated into the surrounding space. During prolonged flight under identical atmospheric conditions the so-called equilibrium skin temperature is established. As the Mach number increases the difference between the regeneration temperature and the equilibrium temperature increases due to an increase in the radiant thermal flux.---the Editor.

when the air temperature in the stratosphere is higher than that in the SA it is necessary to reduce the airspeed of an aircraft made of ordinary aviation materials (aluminum alloys) to the Mach number at which it is possible to maintain the temperature of kinetic heating of its skin below the tolerable values. Thus, when the air temperature is 5° higher than the temperature of the SA the limiting Mach number is reduced to 2.15 by reducing the true airspeed of the aircraft by 7.5 m/sec, which corresponds to approximately 25 km/hr (Figure 4.1). The reason for this is the fact that for aircraft such as the "Concorde" which is constructed of aluminum alloys, the boundary layer temperature (according to French data) should not exceed 180°C. According to preliminary calculations the temperature of the surface of various parts of the Concorde during flight under SA conditions at a speed corresponding to Mach = 2.2 can reach 153° on the nose of the aircraft, 130° on the leading edge of the wing and 119-116° on the surface of the fuselage [19].

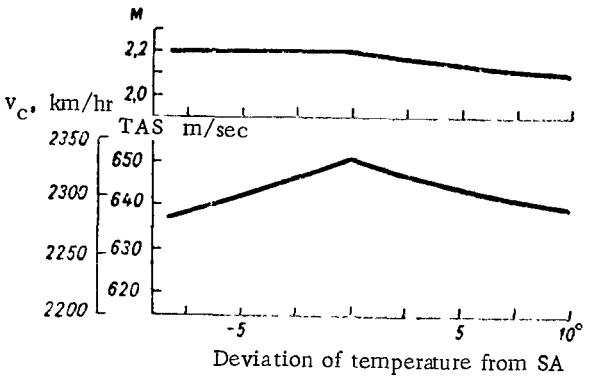


Figure 4.1. Change in Cruise Speed of Aircraft, v_c, Characterized by M = 2.2 as Temperature Deviates from SA Temperature [23]. TAS = True airspeed.

The above temperatures of the surfaces of various parts of the aircraft were obtained for a certain airspeed and given air temperature. However, considerable increases in the velocity of the head-on component of the wind and temperature increases that are common in the real atmosphere in the airlanes will result in additional heating of the leading surfaces of the aircraft. /82

The additional heating of the surface for the above-stated reasons can be calculated using an equation derived as a result of differentiation of equation (4.5), of the following form:

$$\delta T_k = 0.4\ TM\ \delta\ M + (1+ 0.2\ M^2)\delta T. \qquad (4.6)$$

The first term on the right-hand side of this equation determines the kinetic heating of aircraft due to change in the Mach number by δM, caused by an increase in airspeed under the influence of a sudden increment of the velocity of the oncoming component of the wind. The second term determines heating as a result of an increase in the air temperature by δT.

Only if the velocity of the wind in the direction of travel of the aircraft changes and the temperature of the ambient air remains constant, $\delta T = 0$ and the second term of the right-hand side of the equation (4.6) becomes zero. Under this condition equation (4.6) is transformed by means of equation (4.3) to the following form:

$$\delta T_k = 0.02\sqrt{T}\,M\delta V, \tag{4.7}$$

or

$$\delta T_k = 0.001\ V\delta V. \tag{4.7a}$$

In the SA above 11 km $T = 217.5°K$. If under these conditions the Mach number $M = 2$, and the airspeed increment[3] $\delta V = 10$ m/sec, then by equation (4.7) we find $\delta T_k = 3°$.

The results of calculations by equation (4.6) yield the following conclusions: sudden increase in the air temperature at the cruising level of a supersonic aircraft by 1° in relation to its value in the SA at $M = 2$ is equivalent to additional kinetic heating of the skin of this aircraft in the case of the same sudden increase in the velocity of the head-on wind component by 3 m/sec.

These results indicate the need to reduce the limiting kinetic heating temperature of the aircraft skin established for SA conditions by the magnitude of the probable additional heating due to sudden substantial increases in the head-on component of the wind and air temperature pulsations.

[3]The author used an intentionally overstated δV. In the stratosphere such airspeed increments due to the sudden intensification of the head-on component of wind velocity are improbable.---the Editor.

According to preliminary data of the Aviation Technology Service of France the permissible kinetic heating temperature for the Concorde is 127°. Therefore, for flight at altitudes exceeding 11 km this corresponds to M = 2.05 in /83
the standard atmosphere, M = 1.95 in an atmosphere 10° warmer than standard. In an atmosphere 16° warmer than the standard atmosphere the Mach number is correspondingly reduced to 1.9 [24].

The cruising altitude of a supersonic aircraft depends on its cruising velocity and air temperature in the stratosphere. The normal cruise altitude is established for SA conditions. A rise in the air temperature in relation to its value in SA, as stated above, causes a reduction of the thrust of jet engines, with the result that the cruising speed of the aircraft, along with the cruising altitude, is reduced.

If the temperature rises 10 to 15° in comparison to its value in the SA stratosphere the altitude of a supersonic aircraft at M = 2 drops 1 to 1.5 km [2].

Reduction of temperature in relation to the standard atmosphere causes the opposite effect: increase in engine thrust, maximum flight velocity and altitude.

The cruise speed, which determines the maximum flight altitude, is limited by the tolerable temperature of kinetic heating of the skin of the supersonic aircraft.

Limitation of the cruise speed of the Anglo-French supersonic aircraft Concorde to M = 0.25 at standard temperature in the stratosphere permits it to reach an altitude of only about 18,000 m. This aircraft, at the beginning of the cruising phase, will be at an altitude of about 16,000 m. Then it should continue cruising flight at maximum tolerable velocity with gradual ascent for the entire trip to the point of initial deceleration. The speed of the Concorde may correspond to the maximum Mach number M = 2.2 at a temperature 13.4° lower than the SA temperature. Under these conditions it may reach a maximum altitude of 18,900 m [22].

The Soviet supersonic passenger plane, Tu-144 has low frontal drag and high aerodynamic characteristics. Reduction of drag is achieved by reducing the cross-section of the fuselage and by giving it considerably elongated nose and tail parts. The high aerodynamic properties permit the use of a triangular wing with a narrow "beading" jutting forward in relation to it, which lies adjacent to the fuselage. The trailing part of the wing, which has the form of a triangle with rounded edges, is suitable for subsonic speeds. At high supersonic speeds the narrow protruding acute-angular part of the wing is more suitable [12]. In the interest of improving aerodynamic characteristics the aircraft was designed without horizontal tail wings (by the "tailless" scheme). The function of the elevators is performed by the so-called elevons, which combine the functions of the elevators and ailerons.

During supersonic flight the aircraft surface will be heated to 150°. In this connection the design of the aircraft incorporates new aluminum alloys /84 of increased strength and thermal resistance. The engine housings, elevons and rudders, which operate under particularly severe conditions, are made of titanium.

The above-described qualities of the Tu-144 enable it to develop a cruise speed of 2,500 km/hr, which under conditions of the standard atmosphere corresponds to a Mach number of 2.35. The Tu-144 will fly at 20,000 m and will have a maximum range of 6,500 km [1, 3]. For more details concerning the features of the first Soviet supersonic passenger liner the reader is referred to articles published in Soviet journals [3, 8, 11, 12].

4.3. Fuel Consumption as Function of Temperature and Wind

The fuel consumption of turbojet engines (TRD) [*Turboreaktivniy Dvigatel'*] per unit distance is an important economic index of transport aircraft. It depends on atmospheric pressure, air temperature, wind and flight speed.

During flight on a constant level according to the barometric altimeter at a given Mach number the fuel consumption per km is determined only by the air temperature and wind. Here the consumption of fuel by supersonic aircraft will depend more on the ambient air temperature than on the velocity of the wind components in the direction of flight. Thus, for instance, a 1° rise

in temperature over its SA value, according to data of the International Air Traffic Association (IATA), is equivalent to a 2 m/sec increase in the head-on wind component for the Anglo-French Concorde and a 4 m/sec increase for the American Boeing-2707-200 [21].

The fuel consumption norms of TRD are established for SA conditions. An unforeseen temperature rise in relation to the SA value in the near sonic acceleration zone (transition from the speed corresponding to the Mach numbers from 0.95 to 1.6) when it is necessary to increase the rpm (afterburner mode) of the engines, rapidly exhausts fuel reserves.

If a supersonic aircraft, during the period of acceleration and transition through the velocity corresponding to M = 1.0, unexpectedly enters a zone with an air temperature exceeding the SA value by 20° or more, then, in the opinion of the International Civil Aviation Organization (ICAO), under certain circumstances it will be necessary to terminate the flight due to excessive fuel consumption [22]. For the Concorde, for example, the additional fuel reserve in the case of the above conditions will constitute a large percentage of the payload, which is limited to 12,700 kg at normal fuel consumption to about 59,000 kg during a three hour flight.

Even more important, according to preliminary data of ICAO, is the specific fuel consumption when the engines are operating in the afterburner mode in /85
order to accelerate the aircraft to supersonic velocity. A heavily laden Concorde with the engines in the afterburner mode at an altitude of 9,500 m at a speed corresponding to M = 1.0 and at temperature 10° higher than the SA temperature will consume about 31,000 kg of fuel per hour, i.e., one-third more than the normal hourly reserve.

British Overseas Airways Corporation established in 1965 that for supersonic aircraft with a cruising speed corresponding to M = 2.2, 5° temperature rise of the air surrounding it will result in 25% reduction in acceleration in the range from Mach 1.1 to Mach 2.0. Consequently the time and distance required for reaching cruise altitude are increased, and this involves increased fuel consumption [23].

For aircraft whose cruising speed corresponds to Mach 3, the near sonic phase and acceleration phase are the most critical part of the flight profile in relation to fuel consumption. The most probable effect of anomalously high temperature on fuel consumption of a supersonic aircraft whose cruising speed corresponds to Mach 3 is illustrated in Figure 4.2. A temperature 16° higher than standard over the entire flight distance of 5,900 km will lead to an additional fuel consumption of 5,000 kg, in addition to the 82,575 kg required for normal conditions (in SA). About 3,600 kg of the additional fuel will be consumed in the short segment of the flight, amounting to 370 km, in the near sonic ascent phase, while another 1,200 kg will be used for covering a distance of 4,800 km in the cruise phase of the flight. The remaining 200 kg of additional fuel will be used for the final segment of the flight in the deceleration and landing phase. If the above-mentioned deviation of temperature from standard exists only in the near sonic ascent corridor, the dispatcher may safely reduce the number of passengers or cargo by 3,600 kg. In the above--described case the reduction in the economy of the flight would be justified by the safety provided by the required fuel reserve. /86

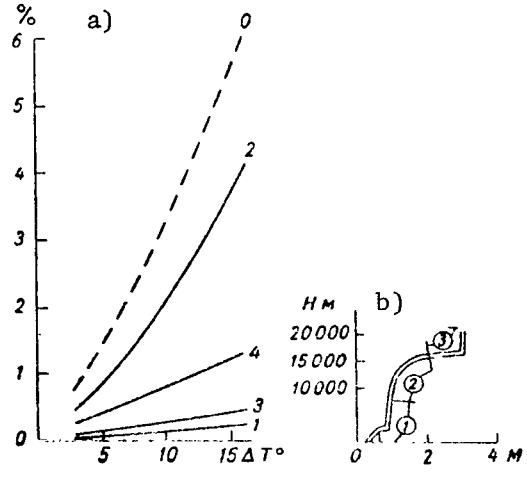


Figure 4.2. Effect of Temperature Changes in Atmosphere on Fuel Consumption in Various Segments of Flight Profile of Brige Aircraft Assuming Constant Weight, Range of 5,900 km and Mach 3 at Cruising Altitude (Nelms, 1968)[23]. a, Increased fuel consumption in % when temperature exceeds standard by ΔT in subsonic ascent phase (1), in near sonic phase (2) in cruise phase (3), in descent and landing phase (4), for the entire flight profile (0); b, Mach number (double line) as function of phases (1, 2, 3) and flight altitude H.

During cruise flight of a supersonic aircraft the temperature usually has little effect on the fuel consumption per km.

When the Mach number and the cruise altitude according to the barometric altimeter are constant the per kilometer fuel consumption remains constant in spite of changes in the air temperature along the way [2]. Under the above conditions the flight range is practically the same as at the SA temperature. If, however, the flight altitude changes as the air temperature rises and falls the per kilometer fuel consumption will change in rather broad limits, whether or not the Mach number is maintained constant. It will drop as the flight altitude is increased and will increase as the flight altitude is decreased.

At altitudes of 18-20 km the wind will have little effect on the flight time of a supersonic aircraft possessing a high cruising speed.

During the winter, as pointed out in Chapter 2, westerly wind at altitudes of 18-20 km usually has lower velocities than in the lower and higher layers, and in the summer there are weak breezes that are directionally unstable over the eastern quarter.

During winter the mean velocity of the west component of a wind of about 25 m/sec (90 km/hr) is quite large over the entire Moscow-Khabarovsk airlane at 20 km. Even this rather rarely encountered velocity, however, will result in only a 5 to 7 minute increase in the flight time of a supersonic aircraft from east to west.

For the above reason the additional fuel consumption in a headwind and a fuel economy in a tailwind will be small.

According to preliminary data of the Aviation Technology Service of France [24], during a flight of the Concorde from Paris to New York (take-off weight 171 tons) the headwind at a velocity of 25 m/sec for the entire trip will result in about the same increase in over-all fuel consumption as a temperature in /87
the atmosphere 10° higher than standard. With the stated temperature rise the additional fuel consumption should amount to about 2,700 kg, or approximately 4% of the total fuel consumption. According to the same data for the Boeing--707-328A, with a maximum take-off weight of 141.4 tons, the increase in the total fuel consumption caused by a headwind with a velocity of 25 m/sec, in the case of flight in the same airlane, is approximately 4,500 kg, or equivalent to 9% of the total fuel consumption.

During the phase of ascent and acceleration to supersonic speed in an atmosphere warmer than standard the fuel consumption by the turbojet engines of the aircraft is especially high. In the case of the Concorde, for instance, at maximum take-off weight the additional fuel consumption during this period in an atmosphere 10° warmer than standard will be 2,300 kg of the 2,700 kg required for the total flight from Paris to New York under the same conditions.

4.4. Flight Regime as Function of Distribution of Temperature and Wind in Troposphere and Stratosphere

Take-off and landing of a supersonic aircraft depend on the temperature and wind above the runway no more than take-off and landing of existing subsonic aircraft. Only a very high air temperature above the runway in the tropics and subtropics, particularly when the airport is located high above sea level, strongly reduces the thrust of turbojet engines during the take-off of a supersonic aircraft. Consequently, it is necessary to reduce its payload.

The rate of ascent at the beginning of ascent and acceleration of a supersonic aircraft will be greater than that of existing aircraft because of the considerable excess thrust of the engines in the subsonic mode.

At an altitude of approximately 4,500 m the rate of ascent of a supersonic aircraft becomes maximum and for the Concorde it is 30 m/sec in the standard atmosphere and 25 m/sec in an atmosphere 10° warmer than the standard [24].

The effect of the deviation of temperature from its SA value on the rate of ascent to 6,000 m is small. However, it becomes considerable at 9,000 m when the aircraft is fully loaded.

During the phase of supersonic acceleration with ascent from 9,000 to 15,000 m and above (depending on the type of aircraft) the effect of increased temperature in relation to the standard temperature becomes critical. This phase of acceleration is located at a distance of 185 to 1,850 km from the point of take-off, depending on the load of the aircraft and air temperature at flight level.

Above 15,000 m the aircraft will continue to accelerate at a smaller angle of ascent until maximum velocity and cruise altitude are reached. /88

Transition of the aircraft to the supersonic mode takes place in the tropopause if its altitude is greater than 9,000 m and less than 15,000 m.

In the case of temperature inversion in the stratosphere at the level of the tropopause or somewhat above it the air temperature has minimal importance, and this is favorable for acceleration of the aircraft to supersonic velocity. Here, however, in the ascent zone near the tropopause, there should be small vertical and horizontal displacements of the head-on component of the wind vector. A combination of the two stated conditions, favorable for transition of the aircraft to supersonic flight, usually occurs at a distance exceeding 300 to 500 km to the right of the jet stream axis (Figure 4.3). Here the altitude of the tropopause is considerably higher than to the left of the jet stream axis and the level of maximum wind velocity is 1-2 km below the tropopause. Therefore transition to supersonic velocity near the tropopause will occur above the layer of considerable vertical and horizontal wind shears over the maximum wind velocity level. Over the extreme southern USSR, to the right of the subtropical jet stream axis, the altitude of the tropical tropopause exceeds 15,000 m. Therefore transition to supersonic velocity can be made there far from this jet stream in the upper cold troposphere, where the temperature is usually lower than in SA.

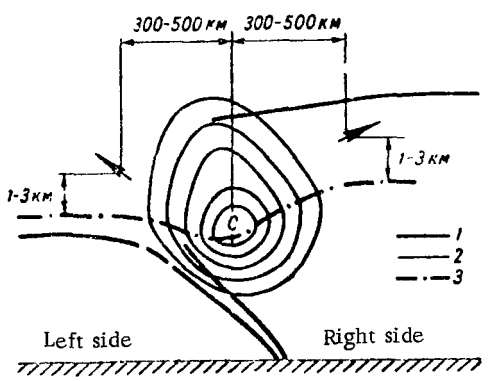


Figure 4.3. Schematic Cross-Section of Jet Stream Normal to Its Axis. 1, Tropopause in front; 2, Isotachy; 3, Maximum wind velocity level; C, Core of jet stream.

The same approach may be used to the tropical zone, since the tropopause there is higher than in the subtropical zone.

If the aircraft must gain altitude to the left of the jet stream axis, where the tropopause is below 9,000 m and the maximum wind level is located above it, acceleration to supersonic velocity should take place in the lower stratosphere 1 to 3 km above the maximum wind level. In the case of isothermy in the

stratosphere, transition to supersonic velocity above the tropopause does not hamper engine operation in the afterburner mode, and even more favorable conditions for this engine operation mode are created in the case of retarded /89 temperature drop with altitude in the stratosphere. Only when there is temperature inversion in the stratosphere will engine thrust above the tropopause be somewhat less than at the minimum temperature level. During the cold part of the year, however, the temperature in the lower stratosphere in the presence of weak inversion is usually lower than in SA. Under such conditions acceleration to supersonic speed in the lower stratosphere does not involve increased fuel consumption.

In the summer, due to the strong warming of the stratosphere in the polar and moderate zones, conditions for transition to supersonic speed are favorable only near the tropopause, where the temperature is lowest. But even the temperature at the level of the tropopause is sometimes higher than in the standard atmosphere. Over Moscow, for example, in July it exceeds -55° in 88% of all cases, and even exceeds -45° in 9% of the cases [4], when transition to supersonic speed becomes unfavorable. Over Dixon Island, located in the polar zone, the temperature at the tropopause level, according to published data [4], is higher than -55° in 96 to 98% of the cases from May through August and higher than -40° in 13 to 14% of the cases in June and July. When the temperature is this high acceleration to supersonic speed becomes difficult.

Over Sverdlovsk the recurrence in July of tropopause temperatures above -55° and above -45° is 56% and 5%, respectively [6], i.e., considerably less than over Moscow, even though the geographical latitude of these two points is about the same (approximately 1° difference). This difference in recurrence of the above-stated temperature limits of the tropopause can be attributed partially to the fact that the data given for Moscow were derived from radio sounding data for the period of 1950-1956, and for Sverdlovsk, for the period of 1959-1963. Moreover, the procedure of determining the tropopause for aeroclimatological purposes has changed since January 1, 1960 [6]. Therefore data concerning the recurrence of tropopause temperature over Sverdlovsk can be regarded as more indicative of conditions of transition to

supersonic speed in the middle latitudes during midsummer than over Moscow. It is noteworthy that in July the tropopause over Sverdlovsk occurs at an altitude from 10 to 13 km in 89% of the cases. These altitudes are quite suitable for beginning transition to supersonic speed, assuming that the maximum wind velocity level is located at least 1 km lower.

The situation over the southern USSR in the summertime is somewhat different. Above Khabarovsk (south of the 50° North latitude), for example, according to data of the Scientific Research Institute of Aeroclimatology (1969), in July tropopause temperature below -55° has a recurrence of about 51%. When the temperature in the tropopause is this low its altitude in the summertime is often greater than 12 km. Consequently the conditions for transition to supersonic speed near the tropopause will be more favorable near this far east point more often than in the Moscow and Sverdlovsk regions. The maximum wind velocity over Khabarovsk, as mentioned in Chapter 3, in most cases did not occur above 13.5 km. In July the middle latitude and polar tropopause over Tashkent is located at altitudes of 8 to 12 km in 47% of the cases, and in 48% of the cases the tropical tropopause is located at altitudes of 15-19 km [6]. The tropopause occurs very rarely in the 12-15 km range, since the axis of the subtropical jet stream is located here. The temperature of the high tropical tropopause over this jet stream, and to the right of its axis, is usually below -55° in July. /90

In the summer the tropical tropopause over the eastern Pamir is located at altitudes of 14-20 km, and usually from 16 to 17.5 km [15]. The temperature at the level of the tropical tropopause varies during this time from -56 to -76°, and the greatest recurrence is in the -64 to -70° interval. The minimum temperature level has the greatest recurrence in the 17 to 19 km layer. In 26% of the cases it coincides with the altitude of the tropical tropopause or is located 250 m above it, and in the other cases, 1 to 3 km above it.

The maximum wind velocity level is established between the polar and tropical tropopause, at altitudes from 11 to 16 km, but usually in the 11 to 12 km interval.

On the basis of the data given for Tashkent and the eastern Pamir it can be concluded that in the summer over central Asia favorable conditions exist for beginning acceleration to supersonic velocity above 14 km.

At lower altitudes ascent in the supersonic mode is not recommended in the zone of strong jet streams with an axial velocity of 100 m/sec or more, since involved here are vertical wind shears exceeding 30 m/sec 1 km below the maximum wind level in 60% of the cases and 1 km above it in 74% of the cases [5].

During summer the low tropopause and stratosphere are considerably warmer in the polar region than the standard atmosphere and will be most unfavorable for ascent of the supersonic transport, particularly during acceleration to supersonic speed.

As indicated above, supersonic aircraft should cruise in the stratosphere, when it is warmer than the standard atmosphere, at limited speed in accordance with the magnitude of difference of the temperatures in the real and standard atmospheres. The actual air temperature at cruising flight altitude can be determined with a sufficient degree of accuracy according to the map of absolute topography of the 50 mb isobaric level (AT_{50}). In the northern hemisphere the mean monthly temperature at the 50 mb level from April through September over the polar region and in the moderate latitudes to 55° /91 is much warmer than standard [13]. In May, June and July it is higher than -40° at 85° latitude, i.e., positive deviation of more than 16° from SA. During individual days the temperature at the 50 mb level over the polar region approaches 35°. Thus, for example, at 3 A.M. Moscow time on May 30, 1968, the air temperature at the 50 mb level over Franz Josefland, northern Novaya Zemlya and the Kara Sea, rose to -34, -36°. Such high temperatures at the cruise altitude make it more economical to fly supersonic aircraft over the North Pole during the summer months. Some limitation on the cruising speed of supersonic aircraft is also required from May through August in the middle latitudes, where the mean monthly temperature at the 50 mb level is 8 to 10° higher than in the SA.

In the summer (April through September) supersonic flights are the most economical in the subtropical and tropical zones, because there the temperature at the 50 mb level, as in the winter, stays lower than in SA.

In the winter (November through February) the mean monthly temperature at the 50 mb level is highest in the middle latitudes, i.e., the 45-55° north latitude zone, and falls toward the pole and equator [13]. The temperature in the specified zone of middle latitudes is 5 to 11° higher than in SA, but is below the SA temperature over the remainder of the northern hemisphere.

During the transition months (including March and October), mean monthly temperature at the 50 mb level over the polar region is about the same as in the SA, and is warmer (6-11° warmer than in SA) in the moderate zone between 55 and 65° north latitude.

The highest temperature in the stratosphere October through March is noted over the Pacific Ocean in the Aleutian anticyclone, and over other parts of the earth during sudden stratospheric warmings. Along the periphery of the Aleutian stratospheric anticyclone above Kamchatka and the lower course of the Amur River the temperature at the 50 mb level quite often is 10 to 15° higher than in SA. During winter stratospheric warmings the temperature at the 50 mb level can be 16 to 32° warmer than in SA (Table 4.1). Supersonic flight at normal cruising speed is impossible in the vicinity of such anomalously high temperature.

During the second half of December, 1967, in the Far East and over southeastern Siberia there was a strong stratospheric warming, described in Chapter 2. It is important to note here that as it spread from east to west it caused an increase in the temperature at the 50 mb level from 17 to 22 December over northern Sakhalin and the Amur Valley to -37°. Toward the end of this period the temperature at the same level over Khabarovsk reached -38° (Figure 4.4). /92
During these days the speed of a supersonic aircraft traveling east from Lake Baykal would have had to have been reduced considerably due to the very high temperature at the cruise altitude. During the last days of December, 1967, there was a second stratospheric warming, which spread from west to east and covered much of Europe. Because of it the temperature at the 50 mb level

TABLE 4.1. CERTAIN EXCEPTIONALLY HIGH TEMPERATURES IN HIGH LATITUDES DURING WINTER WARMINGS [23].

Year	Month	Temperature (°C) at isobaric levels	
		50 mb	30 mb
1957	II	-24	-15
1958	I	-35	-35
1960	I	-35	-28
	II	-35	--
1963	I	-28	-19
1964	I	-31	-24
1965	I	-40	--
	II	-40	--
	III	--	-16
Standard atmosphere		-56	-56

rose on the 30th of December over northwestern and central European USSR to -43, and -49°, respectively. Here a temperature higher than in SA was noted to the Urals and the Aral Sea.

A graphic representation of the thermal conditions at cruising altitude of the supersonic transport Tu-144 over the Soviet Union can be obtained from the temperature distribution at the 50 mb level in the latitudinal and meriodional directions for January, April, July and October, discussed in Chapter 2 (see Tables 2.2 and 2.3).

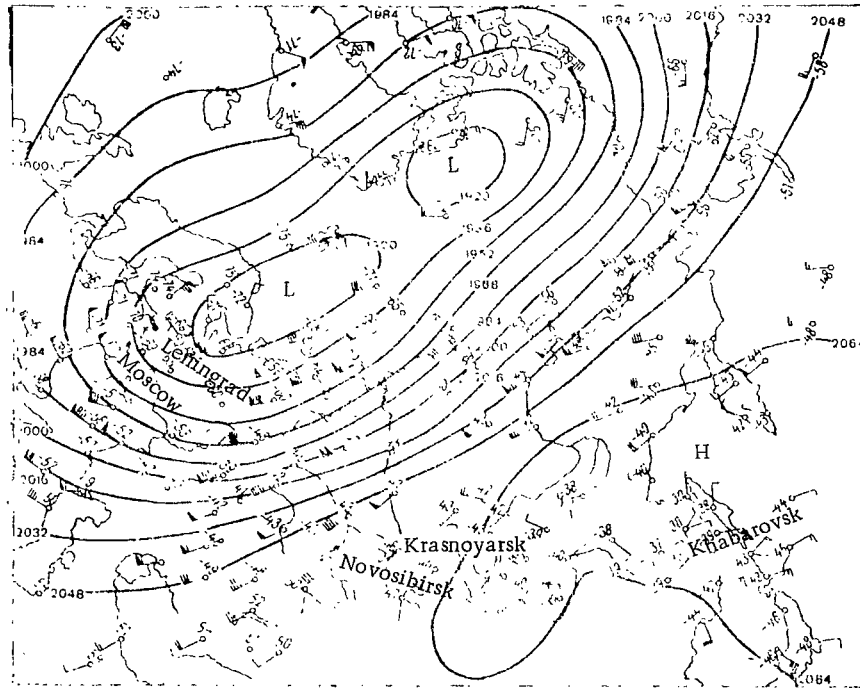


Figure 4.4. AT_{50} Map of 3 A.M., 22 December 1967.

Temperature recurrence above -56° at the 50 mb level is very important in determining the engine operating mode of a supersonic aircraft at cruise altitude. Such data have been published [25]. They showed that the recurrence of temperatures above -56° over southwestern Europe and the USA is less than 50%, but exceeds 95% over the northwestern Pacific Ocean. Temperatures above -46° occur more often than 1/3 of the year over the polar region, but are never encountered south of 40° north latitude.

4.5. Sonic Boom and Its Dependence on Vertical Distribution of Temperature and Wind /93

Sonic shock (boom) is caused by the sudden compression of air near the nose of an aircraft, the velocity of which exceeds the speed of sound. This phenomenon is explained by the laws of gas dynamics [7, 18]. The character of disturbances formed in the air by an aircraft traveling in it depends on the ratio of the speed of the aircraft to the speed of sound, i.e., on the Mach number. An aircraft traveling at subsonic speed stimulates weak disturbances (elastic waves) in the air, which propagate in all directions at the speed of sound. Since the velocity at which the disturbances propagate exceeds the speed of the aircraft, they outdistance the aircraft and encompass the air space in front of it. Under this condition the pressure at the leading surface of the aircraft cannot build up to any great degree in front of the aircraft due to the elasticity of the atmosphere, which rushes from it at the speed of sound. If, however, the aircraft travels at supersonic velocity the disburbances created by it cannot propagate ahead. They are formed behind the nose of the aircraft /94 and are concentrated in a volume of air shaped like a cone (Figure 4.5). The vertex of the cone coincides with the nose of the aircraft, and its axis with the direction of travel of the aircraft. The angle between the generatrices of the conical surface and the trajectory of the aircraft is determined from the relation μ = arc sin 1/M. It is called the angle of weak disturbances or the Mach angle. The magnitude of this angle decreases as the velocity of the aircraft increases.

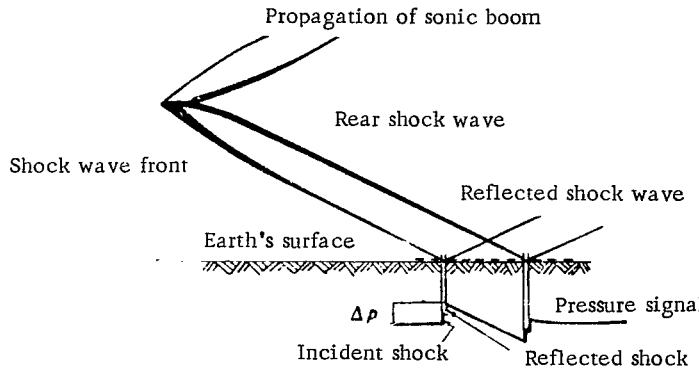


Figure 4.5. Nature of Recording of Pressure Change Resulting from Sonic Boom (British Aircraft Corporation [23]).

At supersonic velocity the air near the leading surfaces of the aircraft undergoes considerable adiabatic compression. Its density and drag increase. Consequently a "compression jump" or shock wave is formed, which is perceived as a sonic shock or sonic boom. Hence the concept of the sound barrier that has to be overcome by an aircraft as its velocity is increased to supersonic.

The point of sudden increase in air density is called a shock wave. The shock wave occurs on the surface of the cone of disturbances and forms the shock wave front.

The shock wave propagates in the direction normal to the generatrices of the cone in the direction of uncompressed air. The speed of the shock wave relative to air depends on the speed of the aircraft and on the difference (drop) of pressure and temperature in the undisturbed air and directly behind the shock wave [7]. Near the origin of the shock wave its velocity exceeds the speed of sound. With distance from the origin, however, the shock wave slows down and travels close to sonic velocity, i.e., is converted into an acoustical (sound) wave [16, 20].

An observer, standing on the earth's surface will hear a sonic boom only when it is confined within the disturbance cone (shock wave front), i.e., sometime after the supersonic aircraft has flown over him. /95

The strength of the sonic boom (sonic shock) on the earth's surface weakens as flight altitude increases. It is a function of the speed of the aircraft, amplitude of the air pressure jump in the near earth layer upon the arrival of the shock wave and of such parameters as the pressure gradient, vertical temperature gradient, vertical wind shear, it also depends to a lesser degree on humidity, cloud cover and hydrometeors [23].

The microbarographic recording of the pressure change associated with the appearance of a shock wave has the shape of the Latin letter N: the sudden pressure rise is followed by a slow drop, and then by rapid return to the former pressure (see Figure 4.5). This pressure jump, generated by a shock wave under SA conditions, travels beneath the trajectory of the supersonic aircraft in the direction of its travel. Maneuvering of the aircraft or the features of vertical distribution of temperature, wind and turbulence caused by them create various distortions of the N-shaped manifestation of the shock wave. In the case of an aircraft flying at supersonic speed, front and rear (tail) shock waves are formed. If these waves reach the earth's surface with an interval of not less than 0.08 seconds they are perceived as two successive booms [20].

The effect of sonic booms generated by the shock wave of an aircraft traveling at supersonic speed on people and surface objects depends on the pressure drop in the shock wave at the earth's surface. If the latter is less than 1.5 kg/cm^2 the sonic boom will be heard as a distant thunder-like explosion. With pressure drops up to 5 kg/cm^2, the boom resembles a nearby explosion. In cases of pressure drops between 5 and 15 kg/m^2 windows are broken, and when the pressure drop approaches 40 kg/m^2 light structures can be damaged [16].

In cross-section through the longitudinal axis of an aircraft in straight horizontal flight the shock wave front can be straight only in isotropic atmosphere, i.e., under conditions when the air temperature and wind remain constant with altitude and in any direction horizontally. Under actual atmospheric conditions it is distorted, depending on the vertical distribution of temperature and wind.

In the case of uniform temperature reduction in the troposphere with altitude, i.e., normal vertical temperature gradient, the sloping shock wave

front becomes concave. This is a result of the fact that when the temperature in the lower layer of the atmosphere is high the speed of sound increases with the propagation of the shock wave front toward the earth's surface in proportion to the square root of temperature, and angle μ increases correspondingly.

In the case of change in the wind with altitude, i.e., when there is a vertical wind displacement, the shock wave front is distorted. It becomes concave if, as it propagates downward, it enters a layer of weakening headwind, and convex if, below flight altitude, it enters a layer of tailwind which weakens as it approaches the earth's surface [23]. /96

The curvature of the shock wave front caused by change in the wind with altitude is the result of geometrical addition of wind velocity vectors at different altitudes to the vector of the speed of sound [20]. The latter is directed perpendicular to the shock wave front toward undisturbed air.

Changes in the speed and location of flying aircraft cause additional distortions of the shock wave front: 1) the shock wave front becomes concave as the speed of the aircraft and, accordingly, the Mach number increase, and convex as the Mach number decreases; 2) if the aircraft is climbing (gaining altitude), the shock wave front, as it propagates downward from the aircraft, becomes more vertical, and if the aircraft is descending, the slope of the shock wave front to the horizon decreases.

It is useful to have an idea of the combined effect of the vertical distribution of temperature and wind on the shock wave profile under certain characteristic atmospheric conditions and in the various phases of the flight of a supersonic aircraft.

Under conditions of the usual drop in air temperature upward to the tropopause and strengthening of headwind with altitude up to flight level (Figure 4.6) the shock wave front from the aircraft is bent forward, becomes vertical in the lower layer and is separated (isolated) from the earth's surface. Here only a noise similar to a roar will reach the earth's surface. There will be no sonic boom, since the shock wave is stretched perpendicular to the vertical /97

segment of the shock wave front in the horizontal plane. Therefore it will not be noticed on the earth's surface. If the aircraft is gaining altitude the point of isolation of the shock wave front in case of a headwind will be higher above the earth's surface than in horizontal flight, and this precludes the danger of the sonic boom. When a supersonic aircraft begins its descent, the point of isolation of the shock wave front can coincide with the earth's surface. However, during rapid reduction of speed as the aircraft descends the shock wave front will have a tendency to curve outward. This weakens the strength of the sonic boom due to scattering of the energy along the lengthened shock wave front.

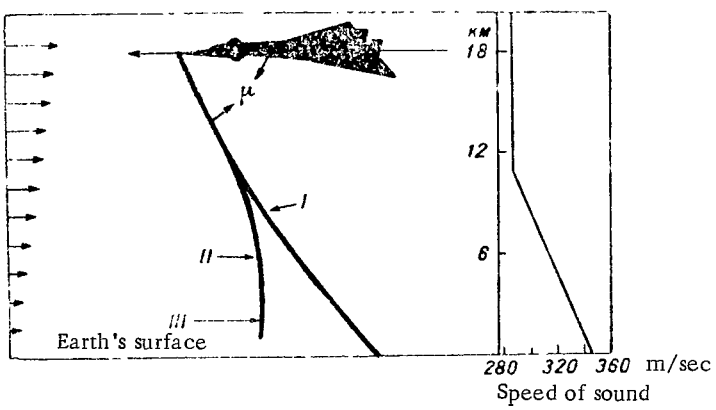


Figure 4.6. Distortion of Shock Wave By Headwind, Strengthening with Altitude, in a Case of Normal Vertical Temperature Distribution [23]. 1, Straight (normal) shock wave in homogeneous atmosphere; II, Shock wave in head wind; III, Sonic boom not heard.

In the case of temperature inversion in the troposphere and tailwind strengthening with altitude (Figure 4.7) the leading edge of the shock wave front has a tendency to bend forward. Here, in view of the law of conservation of energy, the pressure jump will weaken near the earth's surface as it propagates downward along the lengthened shock wave front. Accordingly the sonic boom is also weakened.

It should be noted that changes in the strength of the sonic boom at the earth's surface due to vertical temperature and wind gradients in the atmosphere will be substantial only when the speed of the aircraft becomes greater than the speed of sound at the earth's surface. This corresponds to Mach = 1.3 for flight altitude near the tropopause in a standard atmosphere. During ascent, however, under conditions of a headwind, the strength of the sonic boom may change for values $M \geqslant 1.6$. /98

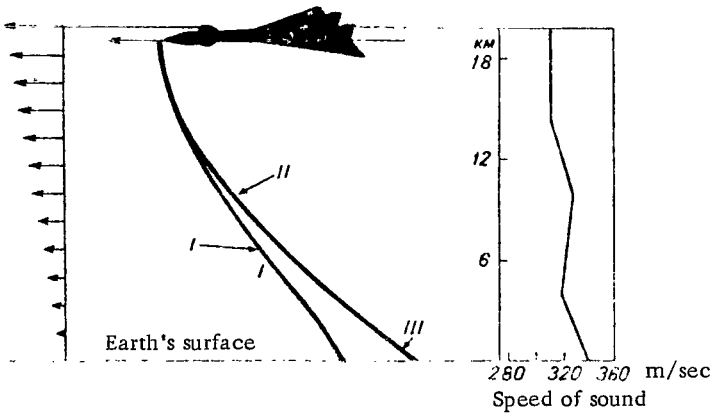
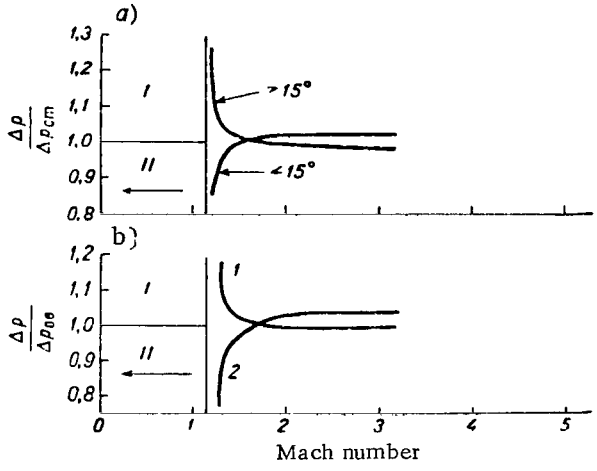


Figure 4.7. Distortion of Shock Wave By Tailwind Strengthening with Altitude and by Temperature Inversion in Bottom Layer [23]. I, Straight (normal) shock wave in homogeneous atmosphere; II, Shock wave in tailwind; III, Sonic boom audible.

Within the limits of the supersonic velocity of horizontal flight at Mach numbers less than 1.3, a headwind and air temperature at the earth's surface higher than in SA may result in a 15% or more increase in the pressure jump compared to its value in motionless SA (Figure 4.8). Under conditions of a tailwind or temperature inversion the pressure jump decreases rapidly at Mach numbers less than 1.3. In cases of flight velocities greater than the limit corresponding to Mach 1.3, the wind and temperature will cause the pressure jump in a sonic boom to change by not more than 5% of its magnitude in the stationary SA [23].



Change in the pressure jump near the earth's surface under the influence of wind will be considerably less if the aircraft at cruise altitude is located much higher than the maximum wind velocity level. Shock waves propagating from the aircraft to the earth's surface through the maximum

Figure 4.8. Effect of Temperature of Earth's Surface (a) and Wind (b) on Pressure Jump Δp, Causing Sonic Boom (Kane and Palmer, (1964), [23]. I, Pressure jump in standard atmosphere; II, Absence of sonic boom at Earth's surface; 1, Headwind; 2, Tailwind.

wind velocity level are, in the general case, greatly weakened. Therefore the probability of damage by sonic boom caused by the shock wave of an aircraft in horizontal flight at cruise altitude without acceleration is undoubtedly diminished.

In general the effect of the headwind on the propagation of the shock wave and on change in the force of the sonic boom is added to the effect of temperature on these phenomena, and the effect of a tailwind is subtracted. The effect of vertical wind displacement is usually relatively minor compared to the effect of vertical temperature gradients [20]. /99

The effect of atmospheric turbulence on the shock wave can heavily distort the form of its microbarographic recording. Consequently, at certain points on the Earth's surface segments of pressure curves during passage of a shock wave front may appear rounded, and at other points there may be distinct peaks. According to data from tests on the propagation of sonic boom over Oklahoma City in the U.S., it was determined that the upper and lower limits of the pressure jump on a shock wave front with a distorted pressure curve are respectively 2.0 and 0.3 times greater compared to the nondistorted recording [23].

As flight altitude increases the sonic boom weakens, because the shock wave spreads through an ever-increasing air volume. According to the law of conservation of energy, as we mentioned above, the pressure jump will weaken by measure of propagation over a larger disturbance cone area. The logarithmic reduction of pressure with increasing flight altitude also leads to a reduction of the pressure jump near the earth's surface [20].

A very heavy aircraft has great lifting force. Therefore it will cause a stronger sonic boom on the earth's surface compared to smaller aircraft flying at the same altitude with lesser lifting force. The amount of lifting force depends also on the type of aircraft. Consequently the most suitable aerodynamic form of aircraft is conducive to a reduction in the strength of the sonic boom.

The effect of flight speed on the strength of the sonic boom increases slowly as the Mach number increases beyond Mach 1.5 [20].

In cruising flight the force of the shock wave and sonic boom should gradually diminish, regardless of meteorological conditions, for two reasons:
1) systematic reduction of the weight of the aircraft with fuel consumption, accompanied by the corresponding reduction of lift of the aircraft;
2) increased limiting flight altitude above sea level.

Supersonic aircraft should break the sound barrier at an altitude of at least 11,000 m above the earth's surface so that the sonic boom will not disturb people and will cause no damage to structures near airports and in the ascent zone [20]. Depending on the performance characteristics of the aircraft and vertical temperature and wind distribution, however, this altitude may be higher.

The strength of the sonic boom at the earth's surface decreases as the angle of ascent of the aircraft increases during the time that the aircraft is climbing. If the angle of ascent is sufficiently large, sonic boom can /100
be avoided. However, the angle of ascent at which there is no sonic boom at the earth's surface increases very rapidly as the Mach number increases: at Mach 1.2 this angle is 3.7°; 14.5° at Mach 1.4 and 30.1° at Mach 2.0 [20].

In horizontal flight under SA conditions at a velocity corresponding to Mach 1.15 sonic boom occurs at the earth's surface only along a line lying directly beneath the trajectory of the aircraft. In the case of smaller Mach numbers the sonic boom does not occur at all. At Mach numbers greater than 1.15 sonic boom is heard not only beneath the trajectory of the aircraft, but also on both sides of it. The width of the band in which the sonic boom is heard on the earth's surface perpendicular to the trajectory of the aircraft, according to theoretical calculations, increases as the Mach number increases. The force of the sonic boom weakens as a function of distance from the projection of the flight path to the earth's surface. On the normal to this line at a distance approximately equal to the flight altitude of the aircraft, sonic boom is sensed by human ears as a roar (Figre 4.9). A cross-wind may cause the pressure jump,

and accordingly the force of the sonic boom, to be greater on one side of the path of the aircraft than directly beneath it.

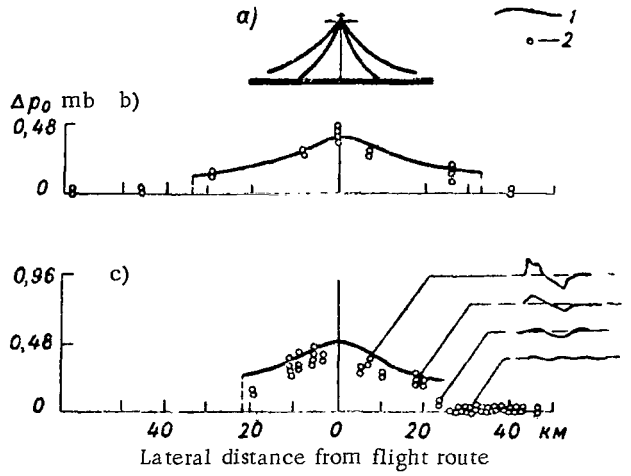


Figure 4.9. Propagation of Sonic Boom in Directions Away from Flight Path [23]. a, Diagram of propagation of shock wave from aircraft; b, Magnitude of pressure jump Δp_0 at Mach 2.0 and flight altitude 15,910 m; c, Pressure jump Δp_0 at M = 1.5 and 11,340 m; 1, Theoretical curve; 2, Observation data.

Propagation of sonic boom /101 in the horizontal direction on both sides of the path of the aircraft also depends on the vertical temperature gradient and wind shear.

It may be necessary to determine the course of the aircraft in the ascent and descent zones, in which the effect of the shock wave at the earth's surface will be minimal. At this time, however, there are not enough experimental data and experience for practical solution of this problem using meteorological information concerning the distribution of the wind and temperature with altitude.

In ensuing years, obviously, meteorological information will be fed into a programming computer, to work out flight plans for supersonic transports and determine the width of the shock wave propagation zone and the magnitude of the pressure jump on the earth's surface.

A program was written in the U. S. in 1965 for forecasting the force of sonic booms with the aid of a computer according to temperature and wind distribution with altitude between the aircraft and the earth's surface [23]. Reflection of the shock waves by the earth's surface and aircraft parameters, such as the Mach number, flight altitude, acceleration, turning radius, lift,

size and weight of the aircraft, were taken into account during development of this program. Predicted values are checked by comparison with the recordings of shock waves caused by supersonic aircraft.

4.6. Requirements on Information Concerning Temperature, Wind and Their Forecasting

Reliable measurements of the temperature in the atmosphere from the Earth's surface up to 25-30 km are required for calculations of speed and altitude of supersonic flight, and also of the fuel consumption of the engines of a supersonic aircraft. This requirement can be satisfied with the corresponding pressure measurement accuracy within this layer of the atmosphere. The existing radio sounding network does not completely satisfy the requirements of supersonic aviation. Research is underway in the Soviet Union and other countries aimed at improving existing or developing new, more accurate radiosondes.

Continuing improvement of the measurement systems of meteorological satellites is approaching the point where it will be possible to obtain the three-dimensional distribution of the temperature and wind fields several times a day in much of the atmosphere.

The plan of the World Meteorological Organization (WMO) for investigation of global atmospheric processes includes provisions for obtaining, with the use of meteorological satellites equipped with multichannel infrared radiometers, the vertical temperature distribution above clouds, including in the stratosphere, /102 with an accuracy up to ±2°C (WMO Bulletin [26]). This plan also includes the use of satellites in synchronous orbits for obtaining data on the wind at various altitudes by photographing the cloud fields in broad zones of the northern and southern hemispheres between 45 and 50°.

It was stated above (§4.5) that the force of sonic boom on the earth's surface depends, in addition to all other conditions, on vertical temperature and wind distribution between the earth's surface and cruising altitude. Consequently, in order to calculate the altitude of transition to supersonic flight in the zone of ascent to crusing altitude, and also in order to determine the altitude of the deceleration to subsonic speed in the descent zone, it is

necessary to have information concerning the vertical profile of temperature and wind from the Earth's surface to the 50 mb isobaric level at least four times a day. This information will also be used for calculating fuel consumption in the ascent and descent zones.

The great influence of considerable positive deviations of air temperatures from its SA value, and also of sudden changes in the wind in the airlane, on the flight mode and fuel consumption give rise to a need for forecasts of temperature and wind distribution on the 50 mb isobaric plane, which is closest to the flight altitude of supersonic transports.

Numerical forecasts of the wind and temperature at isobaric levels from 400 to 200 or 100 mb and also forecasts of tropopause altitude, are presently used in several countries for automatic planning of subsonic flights with the aid of computers. Automatic planning of supersonic transport flights requires additional numerical forecasts of temperature and wind at the 50 mb isobaric level.

In future years computerized numerical forecasts of wind, temperature and tropopause altitude will be used throughout the world. This will be facilitated by the completion of the World Weather Service (WWS) program, in which existing and proposed worldwide and regional meteorological centers, equipped with large computers and equipment for automatic data processing and forecast mapping, are given high priority. *Gidrometsentr SSSR*, [USSR Hydrometeorological Center], Moscow, will perform the functions of one of the three world meteorological centers (WMC's) in the WWS system. The other two WMC's are located in Washington and Melbourne. There will be 20 regional meteorological centers (RMC's) in the USSR "Gidrometsentr" and the meteorological centers in Novosibirsk, Tashkent /103
and Khabarovsk will perform as the RMC's. National meteorological centers will also perform the functions of WMC's and RMC's in other countries.

The International Civil Aviation Organization (ICAO) has established several zone forecast centers based on several national meteorological centers. These centers are responsible for providing forecasts for flights on international airlanes in certain zones. Eventually the development of a system of zone forecasts for aviation will be taken jointly by WMO and ICAO.

In the Soviet Union, forecasts for long-range internal and international airlanes are made by the Moscow Main Aviation Meteorological Center and zonal aviation meteorological centers in Novosibirsk, Tashkent and Khabarovsk. These are national centers. They, in fact, will provide meteorological information and forecasts for supersonic transport flights.

The developing system of centralized meteorological service for long-range flights, based on computer technology and the latest automation methods, will make it possible for synopticians of the aviation meteorological stations at airports to concentrate their efforts on weather forecasts for take-offs and landings, and also for the ascent and airport approach zones.

According to ICAO data [22], changes in the altitude and course of supersonic transports for the purpose of avoiding high temperature zones, large cumulonimbus clouds and turbulence are practically excluded. The reason for this is that the turning radius of an aircraft at supersonic velocity will be extremely large, since the maximum permissible turning rate is 0.3° per second. Therefore sufficiently accurate forecast of temperature and special meteorologican phenomena, and also timely adjustment of them, will be required.

In accordance with the statements made above in §4.4, it will be necessary to avoid unexpected entrance of a supersonic aircraft at maximum cruising speed into a zone of strong sudden warming in the stratosphere. Since these warmings first occur at altitudes of 45-50 km, and then spread downward slowly, it is possible to warn aviation dispatch services of the possibility of high temperature relative to the SA temperature at the flight altitudes of supersonic aircraft. For this purpose aviation meteorological centers should receive telegrams from national and world meteorological centers concerning the beginning of stratospheric warming discovered at points of rocket and radio sounding at high levels (10 mb and above). In future years such information concerning warming in the upper stratosphere and mesosphere will be transmitted by weather satellites. They will also transmit operative information concerning high radiation /104
levels in space in connection with solar flares.

The following facts should be used for forecasting temperature rise at the 50 mb level: stratospheric warming usually weakens as it spreads downward [23].

According to the data in Table 4.1, in six of the seven cases of strong winter warming in the stratosphere observed between 1957 and 1965, the temperature at the 30 mb level was greater than at the 50 mb level. The temperature difference in five cases was 7-9°, and considerably greater in one case. Only in one case was it equal to zero degrees. During the stratospheric warming that occurred in December 1967 over the Far East, the temperature difference between the 30 and 50 mb levels was 2°.

Considerable error in forecasting a temperature higher than in SA in the near sonic acceleration phase may be reflected in reduced fuel reserves for the remainder of the flight, and also may lead to strengthening of the effect of sonic boom on people and surface installations along the airlane [21].

According to IATA data (Table 4.2), the minimum fuel reserve in the tanks of an aircraft depends on the magnitude and sign of deviation of the forecast temperature from the standard value, and also on the magnitude of error of the temperature forecast. During summer, when temperatures higher than the SA temperature predominate in the stratosphere in the high and middle latitudes, the consequences of temperature forecast error will be greatest. On the other hand, during the summer months the day to day fluctuations in temperature at the 50 mb isobaric level are slight and generally do not exceed ±2° [21]. As regards wind, as pointed out in Chapter 3 its velocity at the 50 mb level is usually small, and its direction variable--from west to east through north or south. Therefore, for aeronavigational calculations in the summer it is better to use the initial information concerning temperature and wind at cruising altitude based on double sounding than numerical forecasts of the fields of these elements.

During winter and the transition months (March, October) the day to day variability of temperature and wind in the stratosphere is considerable. Therefore, along with initial temperature and wind data for the 50 mb isobaric level, it is also desirable to have forecasts of these elements for that level. /105

TABLE 4.2. DEPENDENCE OF MINIMAL FUEL RESERVES ON TEMPERATURE FORECASTS [21].

Deviation of forecast temp. from SA temps. in degrees	Minimal fuel reserves (kg) for temperature forecast errors, in degrees			
	2	4	2	4
	Concorde		Boeing-2707-200	
-5	Insignificant	140	450	950
0	270	600	900	1,800
5	340	750	1,000	2,100

For flights of subsonic jet aircraft on international airlanes the following requirements are presently imposed on forecast accuracy for the routes: permissible error in air temperature forecast ±3°, vector difference between forecast and actual wind ±8 m/sec [10]. It is pointed out in an IATA (International Air Traffic Association) document that if fuel reserves in the flight were affected only by air temperature it would be necessary to increase its forecast accuracy to 2°. Since, however, there are many other factors that influence the total amount of reserve fuel, they must be compared with each other in order to determine the relative importance of each.

The amount of reserve fuel required to compensate for temperature forecast error in 85% of all flights over the North Atlantic (from Europe to North America and back) for one year is shown in Table 4.3. The normal law of temperature distribution at flight altitude was used for the calculations. In the case of an arithmetic mean square error not exceeding 3° the required fuel reserve should be equal to 430 and 1,220 kg, respectively. Taken together, all other factors (such as air travel service limitations relative to flight altitude and route, delays at points of departure, etc.) require considerably greater fuel reserves. Thus, the present temperature forecast accuracy is sufficient for calculations of the total fuel reserves for supersonic aircraft.

TABLE 4.3. AMOUNT OF FUEL RESERVES FOR COMPENSATION OF TEMPERATURE FORECAST ERROR FOR NORTH ATLANTIC FLIGHT [21].

Type of aircraft	Fuel reserves in kg
Concorde	300
Boeing-2707-200	800

It is known that if flights over the North Atlantic are planned with consideration of the monthly air temperature, the arithmetic mean deviation over the year will be about 2°. It will be somewhat greater in the winter and smaller in the summer. Therefore a temperature forecast is required in cases of relatively rapid and substantial changes, particularly during the winter [21]. Consequently there is a need for forecasting considerable temperature changes along the flight route, even with the presently acceptable error of ±3°.

Forecasting of substantial wind changes with the presently accepted accuracy is also permissible, since according to the data presented above for the Concorde, an 8 m/sec error in tailwind velocity corresponds approximately /106 to a temperature forecast error of 4°.

During error navigational calculation it is important to consider the combined effect of air temperature and wind. For these purposes the concept of equivalent wind, which takes into account the air temperature, can be used.

4.7. Forms of Presentation of Meteorological Information for Flight Planning

Up until the present time aviation placed requirements on the presentation of meteorological information for planning and execution of flights chiefly in graphic form: forecast maps of standard isobaric levels with wind and temperature data; tropopause maps and maps of vertical wind shear or tropopause and maximum wind; surface forecast maps, giving an idea of the baric field and atmospheric fronts; and forecast maps of special weather phenomena at flight altitude. In individual cases graphic presentation of forecast of flight conditions along the airlanes in the form of vertical cross-sections was required. In addition to the graphic form of information, airlane forecasts in the form of a text, which a pilot can understand without additional explanation, and also tables with wind and temperature forecasts in digital form for segments of a route, are used.

During recent years the successful development of automatic flight planning using electronic computers increased the requirement for digital meteorological data [9]. The possibilities of computerized automated flight meteorological service are considered in [10]. Wind and temperature forecasts, made with the aid of the computer at meteorological centers for specific points (centers) of a regular geographical grid, for fixed levels, fixed time, are presented at the present time as digital information. This information is used for flight planning with the aid of specialized computers, as well as air traffic service in the air space.

It should be pointed out that despite the progressive use of digital information, there is no letup in the need for information from zonal forecast centers in graphic form as prognostic maps, transmitted over facsimile communication channels and used for preflight planning, where this is done without the aid of the computer, and also for compilation of flight documentation given to airline crews.

The existing system of providing aviation meteorological data in the form of tables and maps will presumably continue to operate in many regions of the earth. In the near future, however, digital meteorological information will /107
entirely supplant the tables and prognostic maps. This will happen when electronic planning computers, in addition to wind and temperature, will be fed forecasts of the boundaries of cloud fields in the horizontal plant and vertically; particularly the dimensions of the zones of development of cumulonimbus clouds and the altitude of their bases and crests, and also associated zones of thunderstorm activity, hail, rain, and thermal turbulence of varying intensity, horizontal and vertical dimensions of zones od dynamic clear air turbulence with an indication of the quantitative indices of its intensity.

Digital meteorological information for automatic planning of the flights of supersonic aircraft at airports of departure, and also in the air aboard these aircraft, equipped with special computers will assume particular importance.

In connection with the great importance of digital meteorological information, the extraordinary session of the Commission of Aviation Meteorology, WMO,

jointly with the Sixth Air Navigational Conference of ICAO adopted, in May, 1969, the following recommendations: 1) that ICAO develop aviation requirements for providing and presenting processed meteorological data in digital form; 2) that WMO take these requirements under advisement and examine the possibility of standardizing the system of geographical networks and the content of digital information, codes for its transmission, issuance frequency and effective time of forecast required for regional and interregional exchange of processed meteorological data in the World Weather Service system.

CHAPTER 5 /108

TURBULENCE IN STRATOSPHERE

5.1. State of the Problem

Atmospheric turbulence, which causes buffeting of aircraft during flight, is one of several weather phenomena that pose a hazard to aviation, particularly in cases of clear air turbulence. Extensive research conducted during the last 10-15 years has yielded an idea of the meteorological conditions favorable for the development of turbulence in the troposphere and lower layers of the stratosphere to an altitude of 12 km.

For lack of reliable observation data, very little is yet known concerning turbulence in the higher layers of the stratosphere. The same conclusions that have been derived concerning turbulence in the upper troposphere obviously cannot always be applied to flights of supersonic aircraft in the stratosphere. The problem is further complicated by the fact that supersonic (SST) operation requires knowledge of the flight conditions not only at cruise altitude (approximately 20 km), but also in ascent and descent zones, as well as in the zone of transition to supersonic velocity, where aircraft buffeting is particularly undesirable.

It was once thought that there was no turbulence capable of causing buffeting at cruise altitudes of SST. However, reports of military pilots and experimental flight investigations that have been conducted recently in the stratosphere [2, 3, 5, 21, 34, 35], revealed that the recurrence of turbulence in the stratosphere is less than the upper troposphere, but it can be encountered nevertheless and is a serious obstacle to flight.

It should be stressed that cases of turbulence encountered in the stratosphere are almost always clear air turbulence, which is particularly hazardous because of its unexpected action on aircraft. In connection with the forthcoming operation of supersonic transport aviation, aviation meteorologists are faced with urgent problems related to analysis of the /109
turbulence in the stratosphere and development of methods of forecasting it.

5.2. The Effect of Turbulence on Flight of Supersonic Aircraft

In straight horizontal flight at constant speed equilibrium is established between aircraft weight G and lift A

$$\frac{A}{G} = n = 1. \tag{5.1}$$

In this case aircraft overload n = 1.

During flight in the turbulent atmosphere the basic parameters of aircraft motion are disrupted due to the effect of random air gusts, and it begins to experience additional overloads. The magnitude of vertical overload of the aircraft in this case can be calculated using the following equation:

$$n = 1 \pm \frac{\rho V W_{ef} C_y^{\alpha}}{2G/S}, \tag{5.2}$$

where ρ is the density of air, V is air speed of the aircraft, W_{ef} is the so-called effective vertical velocity of a gust, C_y^{α} is the derivative of the lift coefficient in terms of the angle of attack, G is aircraft flight weight, S is wing bearing area. The (+) sign pertains to the case of a rising current and the minus sign (-) to the case of a descending current. The second term in equation (5.2) represents overload increment Δn. In meteorology overload increment Δn is often simply called overload.

It is clear from equation (5.2) that the overload increment is directly proportional to the air density, speed of the aircraft, vertical wind gust velocity, lift coefficient derivative for angle of attack and inversely proportional to specific load on the wing. Hence it follows, for example, that as the altitude increases the same wind gusts will cause smaller overloads due to reduction of air densities; toward the end of the flight the aircraft will be more vulnerable to the action of turbulent gusts due to reduction of specific load on the wings; the magnitude of overload, other conditions being equal, will increase as the flight speed is increased.

To ensure flight safety in turbulent atmosphere the limits of permissible flight speeds are established for subsonic aircraft in accordance with aircraft

strength requirements on the one hand, and the possibility of exceeding the critical angles of attack on the other [16].

In view of the fact that the speed of the supersonic aircraft is measured in Mach numbers we will include in equation (5.2) the number M, which is the ratio of the airspeed of the aircraft to the speed of sound, /110

$$M = \frac{V}{a} = \frac{V}{\sqrt{xRT}},$$

where $x = \frac{c_p}{c_v}$ is the ratio of the specific heat capacity at constant pressure to the specific heat capacity at constant volume, R is the gas constant, T is the air temperature on the absolute scale.

After substitution into equation (5.2) we obtain

$$n = 1 \pm \frac{\rho M \sqrt{xRT} W_{ef} C_y^{\alpha}}{2G/S}. \qquad (5.3)$$

The Mach number, as we know, is the similarity criterion with respect to compressibility of air and depends largely on the ambient temperature. It is obvious from equation (5.3) that aircraft overloads can be the direct result of sudden changes in the horizontal temperature field in the stratosphere, in addition to other factors.

Airframe overload is illustrated in Figure 5.1 as a function of the Mach number for two different loads on the wing [36]. The graph shows that reduction of the airspeed decreases the magnitude of the overload more at subsonic than at supersonic velocities.

The intensity of aircraft buffeting as a function of the overload increment can be evaluated using the following scale:

Buffeting Intensity	Overload Increment in Fractions of g	Symbols	
		USSR	ICAO
Weak	±0.2	δ^1	
Moderate	±0.5	δ^2	
Strong	±1.0	δ^3	
Violent	±1.1	δ^4	

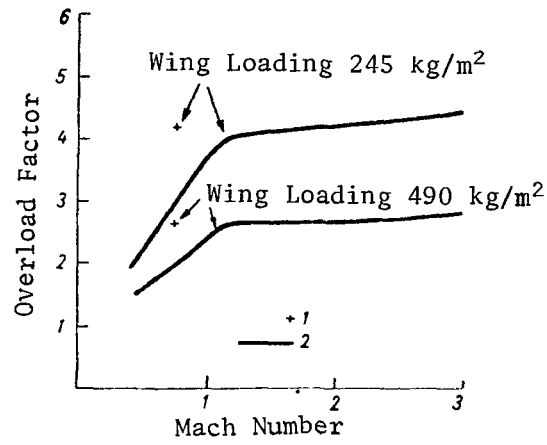


Figure 5.1. Airframe Overload As Function of Mach Number.
1 -- Subsonic aircraft, 2 -- Supersonic aircraft.

It should be pointed out that visual evaluation of the intensity of /111
buffeting of an aircraft by the pilots sometimes does not coincide with the actual AIREPS on aircraft overloads. The pilots often report moderate buffeting as weak, and vice-versa. Therefore, corrections to the AIREP[4] were made at the session of the Commission of Aviation of Meteorology of the WMO jointly with the Sixth Conference on Air Navigation, ICAO, held April-May, 1969, according to which the readings of accelerometers, installed at the aircraft's center of gravity, were changed from 0.5 to 1.0 g for moderate turbulence with slight changes in airspeed, and in excess of 1.0 g for strong turbulence and large changes in the airspeed. Thus, according to the new scale the evaluation of the intensity of turbulence according to the magnitude of overload of the aircraft was shifted by one gradation. Moderate turbulence became known as weak, and strong as moderate.

The scales of turbulent eddies and wavelike disturbances that act on subsonic and supersonic aircraft differ considerably. If subsonic jet aircraft

[4]AIREP is a form for recording meteorological observations during flight.

react most to atmospheric disturbances with dimensions on the order of 200-500 m, then supersonic aircraft will be sensitive to disturbances with dimensions up to several km. According to published data [35], disturbances up to 1,500-4,300 m in length will influence flight at Mach 3. The dimensions of eddies are shown in Figure 5.2 as functions of the spectral density of turbulent energy (which characterizes the intensity of turbulence) for various weather conditions. As expected, the spectral density for all eddy dimensions for clear air turbulence are less than in cumulus clouds or storms. The cross-hatched regions show the ranges of wavelengths that are important for present day subsonic aircraft (Mach 0.8) and supersonic aircraft flying at Mach 3 [35]. We see in Figure 5.2 that the spectrum of disturbances that have an effect on SST falls in the 1-10 km range. Consequently, measurements of /112
turbulence spectra, obtained with the aid of subsonic aircraft, cannot be applied with sufficient reliability to supersonic aircraft.

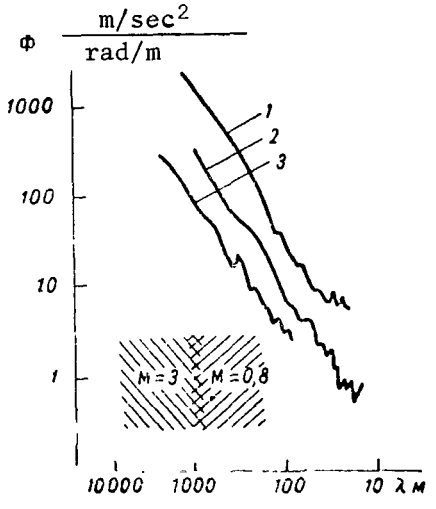


Figure 5.2. Curves of Spectral Density ϕ for Various Weather Conditions.
1 -- Thunderstorm, 2 -- cumulus cloud, 3 -- clear air.

Generally speaking, only the effect of vertical pulsations of wind velocity on aircraft, the influence of which is several times greater than that of horizontal pulsations, is taken into account in analysis of good performance of an aircraft in turbulent atmosphere at subsonic speeds. As pointed out in [20], the contribution of horizontal pulsations of wind velocity to the magnitude of overload of an aircraft at speeds substantially exceeding the speed of sound must be subjected to careful experimental investigation, since the role of horizontal pulsations increases as the airspeed of the aircraft increases. Considering the fact that the magnitude of vertical pulsations in the stratosphere generally decreases in comparison with the troposphere, the difference between the contribution of vertical and horizontal pulsations of various meteorological elements to the magnitude of overload should level off rapidly, and horizontal pulsations of wind velocity and temperature begin to play the same role as vertical gusts.

An aircraft can experience buffeting not only in turbulent atmosphere, but also during flight in laminar wavelike flow. The amount of aircraft overload depends both on the vertical wind velocity at the peaks and valleys of the waves, and on the relationship between wavelength, flight speed and frequency of natural vibrations of the aircraft. The period of accelerations experienced by an aircraft in the wavelike flow will be $\tau \equiv \lambda/V$, where V is the airspeed of the aircraft, λ is the length of the airwave.

When τ approaches the period of natural resonance vibrations of the aircraft the overloads can reach critical values which are dangerous in terms of its structural strength. Supersonic aircraft, obviously, will be more stable to the wave disturbances because the range of waves acting on SST is much broader than for subsonic aircraft.

The effects of buffeting of a supersonic aircraft may also create meso- and micro irregularities in the horizontal temperature field in the atmosphere, as already mentioned above. Assuming that the temperature discontinuity surface in the atmosphere is not smooth, but wavelike, a supersonic aircraft, as it passes through such temperature waves, may experience vibration similar to buffeting in turbulence [30]. This is the result of sudden

changes in the drag of the aircraft during flight in atmosphere that is heterogeneous with respect to temperature. Such buffeting is entirely the result of the temperature structure of the atmosphere and is not related to wind gusts and turbulence. Such conditions presumably are more frequent in /113 the strong temperature inversions in the jet stream zone.

Thus buffeting of supersonic aircraft can be caused both by atmospheric turbulence and by wavelike laminar flow or temperature heterogeneity.

Atmospheric turbulence will have an effect on SST operations in both cruising altitude and during ascent and/or descent and transition from subsonic to supersonic speed (the near sonic phase). Ascent and descent in the troposphere will be made at subsonic speed by supersonic aircraft, and therefore the conditions of buffeting in the troposphere will be the same for it as for subsonic jets.

It is very important that supersonic aircraft avoid the effects of turbulence during transition from subsonic to supersonic speed, since during this phase of flight it is extremely undesirable to subject the aircraft to additional actions that are capable of producing overloads, changing the altitude or speed of the flight. Therefore the near sonic phase of the flight, which will take place near the tropopause, or above it, should avoid cumulo--nimbus clouds and zones of large horizontal and vertical temperature and wind gradients.

It should be pointed out that the intensity of buffeting of supersonic aircraft depends not only on the turbulent state of the atmosphere, but also on their aerodynamic characteristics [20]. It should be kept in mind that in the stratosphere at high M numbers, natural damping of the natural vibrations of aircraft is substantially less, therefore the effects of atmospheric disturbances last for a longer period of time. Thus, a combination of high speed and low air density in the stratosphere will have the effect of considerably lower levels of damping of vibrations than in the troposphere and at subsonic speeds.

5.3. Characteristics of Turbulent Zones in the Stratosphere

There is presently very little information concerning turbulence in the stratosphere obtained by means of aircraft measurements. The result obtained in the US [35] at altitudes up to 23 km by means of a specially equipped subsonic U-2 aircraft are noteworthy (Table 5.1).

80% of the U-2 flights were made over the US and 20% over Western Europe, Turkey and Japan.

/114

TABLE 5.1. PROBABILITY OF TURBULENT ZONES BY ALTITUDE ACCORDING TO AIRCRAFT MEASUREMENT DATA (US).

Altitude, km	Total Flight, km	Flight and Turbulence Zones	
		km	%
6.1-9.1	30,800	2,400	7.8
9.1-12.2	42,800	4,320	10.1
12.2-15.2	79,800	2,000	2.5
15.2-18.3	228,000	4,250	1.9
18.3-21.3	928,000	5,280	0.6
21.3-22.9	11,300	14	0.1

The data presented in Table 5.1 show that the greatest ratio of flight distance in the turbulence zones (10.1%) to the total flight distance corresponds to the 9-12 km interval, i.e., near the tropopause. Further, as altitude increases this ratio decreases, and at altitudes exceeding 15 km it amounts to only 2%. Figure 5.3, in which data obtained on the basis of U-2 flights are represented by the broken line, gives a general idea of the distribution of the recurrence of turbulence with altitude for various geographical regions [5]. The horizontal extent of turbulence zones in the stratosphere (curve 1 in Figure 5.4) can be judged on the basis of these data: in 70% of the cases through flight of the turbulence zones does not exceed 30 km, and only in 0.5% of the cases does it exceed 100 km. The dimensions of turbulence zones are somewhat greater according to data from flights in the jet streams. In particular, zones with a length of 130 km make up about 10% of the total. Turbulent zones are generally somewhat shorter in the stratosphere

than in the troposphere. The average thickness of the turbulent zones varies from 600-1,000 m.

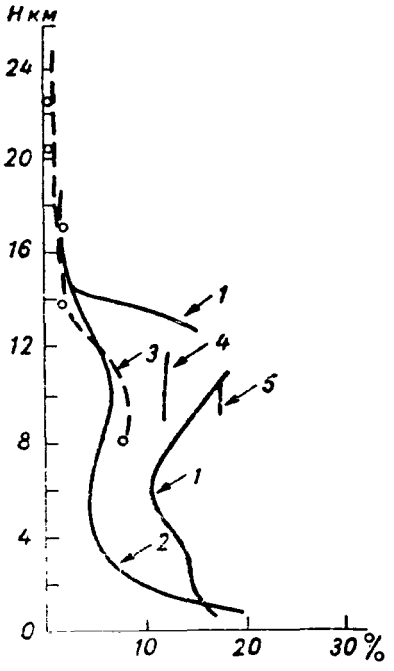


Figure 5.3. Recurrence of Turbulence with Altitude. 1 -- USSR, southern latitudes, 2 -- USA, 3 -- USA, U-2 data, 4 -- USSR, middle latitudes, 5 -- London-Far East and London-North African routes.

Intensity of turbulence, which may be defined as the velocity of the effective vertical gusts, proportional to the overload acting on the aircraft during flight in turbulent atmosphere, decreases as the altitude increases. At altitudes of 18-21 km the velocities of the maximum effective gusts are 2-3 times less, on the average, than at altitudes of 6-12 km. Table 5.2, which provides data on the probability of encountering effective gusts at altitudes of 12-21 km [5], gives the general idea of turbulence intensity. /115 The probability of the encounter is characterized by the distance (in km) that must be flown in order to encounter an effective gust of the given magnitude.

TABLE 5.2. ENCOUNTER OF EFFECTIVE GUSTS OF GIVEN MAGNITUDE AT ALTITUDES OF 12-21 KM IN CLOUDLESS ATMOSPHERE.

Magnitude of Effective Vertical Gusts in m/sec	Distance in km/gust	
	Ordinary Conditions	In Zones of Strong Turbulence
0.6	160	2
1.2	1,600	16
1.8	16,000	160
2.7	160,000	1,600

Table 5.2 clearly shows that as the gusts increase the distance over which it is possible to encounter an effective gust of the given magnitude increases rapidly.

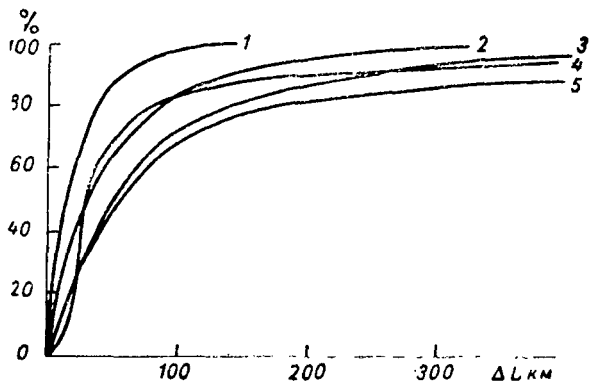


Figure 5.4. Integral Recurrence (%) of Horizontal Extent of Turbulence Zones. 1 -- US (stratosphere), 2 -- US (upper troposphere), 3 -- USSR (middle latitudes), 4 -- Canada, 5 -- USSR (southern latitudes).

The cited data on the intensity of turbulent gusts correspond to certain average conditions. In individual cases gusts of considerably greater magnitude have been recorded. Thus, for instance, during a U-2 flight over Japan at an altitude of 16,800 m the velocity of an effective vertical gust is established at about 6 m/sec.

Turbulent zones of unusual length, 250-300 km, have also been recorded /116 at 16 km over Japan. In such cases the Japanese Islands were under the influence of a typhoon, and the wind velocities in the jet streams at 16 km

reached 50-80 m/sec. This indicates that turbulence in the stratosphere in individual cases can considerably exceed the average values mentioned above.

Pilot observation data of turbulence during flight in the stratosphere at subsonic and supersonic speeds were analyzed in the Soviet Union. This analysis made it possible to establish certain statistical characteristics of turbulence [21]. Analysis of more than 4,000 flights, during 429 of which there was buffeting of various intensities, revealed that the length of turbulence zones in the stratosphere is about 1% of the length of stratospheric flights. One case of buffeting of any intensity is encountered per 7,300 km of flight in the stratosphere. To encounter buffeting of moderate and higher intensity it is necessary to fly a distance of 11,400 km, and to encounter strong and very strong buffeting, 33,000 km. The average recurrence of buffeting (without consideration of intensity) at altitudes of 11,000-20,000 m is 9.6%. Weak and moderate buffeting are most common. The recurrence of strong and very strong buffeting is only 2.2%. The greatest recurrence of buffeting over the middle latitudes of the USSR occurs during the winter months, and the least during the summer.

Data on the dimensions of turbulent zones in the stratosphere were obtained on the basis of these various investigations. Thus, during the period of observations the length of buffeting zones vary from 13-250 km. The maximum recurrence of zone length with strong turbulence corresponds to the 25-100 km gradation, and the average length of buffeting zones is 62 km. The thickness of turbulent zones in the stratosphere does not exceed 300 m in 74% of the cases, although in individual cases it can exceed 1,000 m.

In general these results are close to foreign data obtained from U-2 flights.

The USSR has also conducted special radio sounding observations, which give a qualitative picture of the distribution of recurrence and intensity of turbulence in the stratosphere by seasons [1]. These observations have shown that considerable turbulence, the recurrence of which diminishes with altitude, is often encountered in the stratosphere during all seasons, particularly winter. The maximum recurrence of turbulence occurs in the

winter at altitudes of about 15 km and 18-19 km. The recurrence and intensity of turbulence in the stratosphere are considerably less during summer.

U-2 measurements also show that maximum recurrence of turbulence in the stratosphere occurs during the cold part of the year (January-March). The high recurrence of turbulence during the winter is attributed to stronger /117
atmospheric circulation during this season at the corresponding altitudes.

5.4. Meteorological Conditions Under which Turbulent Zones Develop in the Stratosphere

Since there are practically no clouds in the stratosphere at the proposed flight altitudes of supersonic aircraft (with the exception of the tropical regions, where cumulonimbus clouds can develop to altitudes of 18 km and higher [36]), the turbulent zones encountered here may relate to the clear air variety (CAT -- clear air turbulence).

Investigation of the conditions under which intensive turbulence develops in the troposphere showed that it occurs in those regions where high vertical and horizontal temperature and wind gradients exist, which, in turn, develop in certain synoptic positions [15-17]. To what extent the conclusions that have been derived in the troposphere will be valid for the stratosphere is not yet clearly known. Even now, however, certain features of the structure of the stratosphere and the processes that occur therein can be discussed, and their role in the formation of turbulent zones can be evaluated. We will examine these features.

1. Wind Distribution

Obviously the jet streams in the stratosphere, just as in the troposphere, should have a considerable influence on the development of turbulence.

It was established [15] that there is considerable turbulence in the region of the jet streams in the upper troposphere and the lower stratosphere, causing buffeting of the aircraft. This turbulence is encountered principally in places where there are high horizontal wind gradients (shear) across and along the flow. Turbulence in the presence of lateral wind shears is generally observed on the cyclonic side of the jet stream in the zone of relatively

strong wind. The width of this zone in the direction normal to the flow is 100-150 km.

Of particular importance is the delta of the high-altitude frontal zone (HAFZ -- high altitude frontal zone), where the high-altitude air streams diverge and where, consequently, wind shears, characterized by weakening of the wind in the direction of the flow, develop along the flow. Here the curvature of the isohypses often changes from cyclonic or straight to anticyclonic.

Figure 5.5 shows a unified diagram of the synoptic situation in which are distinguished zones that are most favorable for development of clear air turbulence in the troposphere [17]. The recurrence of turbulence for various positions of the barometric field at flight altitude is given in Table 5.3. /118 The table clearly shows that buffeting occurs in 67% of the cases in the zone of diversions of airflows and near the axis of the ridge.

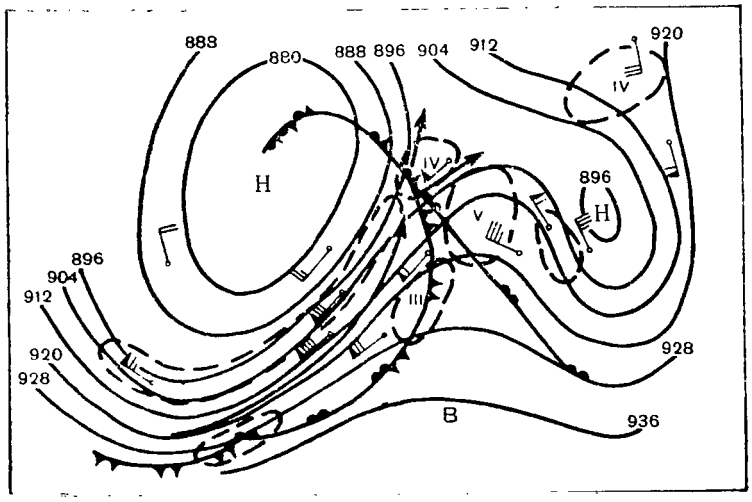


Figure 5.5. Diagram of Synoptic Situation Characterizing the Arrangment of Zones that are Most Favorable for the Development of Clear Air Turbulence. The broken line and Roman numerals denote regions of different synoptic conditions.

TABLE 5.3. RECURRENCE (%) OF MODERATE AND STRONG BUFFETING (180 CASES)

Synoptic Situation	Recurrence of Buffeting in %
Leading Part of Trough:	
Cyclonic Part of Jet Stream	18
Anticyclonic Part of Jet Stream	32
Near Ridge Axis.....	27
Trailing Part of Trough:	
Cyclonic Part of Jet Stream	6
Anticyclonic Part of Jet Stream.....	8
Near Center of Cyclone.....	1
Trough.....	8

TABLE 5.4. RECURRENCE (%) OF HORIZONTAL WIND SHEARS > 2m/sec FOR 100 km

Level in mbar	Horizontal Wind Shear in m/sec for 100 km					Number of Cases
	2.1-2.5	2.6-3.0	3.1-3.5	3.6-4.0	4.1-4.5	
100	64	23	10	3	--	300
50	68	22	8.2	2.4	--	290
30	54.5	24	13	7	1.5	270

The results of divergence of the jet steams in the stratosphere, as mentioned in Chapter 1, are often encountered in the region of the winter stratospheric cyclone and particularly over the northwestern regions of Europe and North America. There is reason to assume that these zones, analogous to the troposphere, can be the centers of turbulent zones in the stratosphere.

The data presented in Table 5.4 give an idea of the magnitudes of horizontal wind shears and their recurrence for various levels of the stratosphere.

The table shows that the recurrence of high horizontal wind gradients in the stratosphere has a tendency to increase with altitude. These data are considered to be the average values for relatively great distances (on the order of 1,000 km), and therefore give only a general picture of the distribution of horizontal gradients in the stratosphere. The lack of measurements of a smaller scale prevents judgement concerning the meso-irregularities of the

horizontal wind field. It is theorized, however, that these elements may be considerable in the stratosphere, especially in the HAFZ region [22].

Recent analysis of [21] cases of aircraft buffeting in the stratosphere confirm the great importance of horizontal wind shears for the formation of turbulence. It has been found that great changes in wind velocity in the stratosphere over a given point with the passage of time, caused by large horizontal wind velocity gradients along and across the flow, are accompanied by decreased recurrence of aircraft buffeting. Thus, when the wind velocity changes by 10 m/sec and more during the last six hours, buffeting occurs in 68% of the cases, and when the wind velocity changes by 20 m/sec and more the recurrence of buffeting of the 6 hour period is 74%.

One of the basic reasons that turbulence develops in the atmosphere is vertical displacement of the wind vector. In studies of the dependence of the probability of buffeting on the magnitude of the vertical wind shear, conducted on the basis of analysis of a large number of reports of buffeting in the upper troposphere [4], have shown that the probability of buffeting increases very unevenly as the displacement of the wind vector increases. The maximum probability of buffeting at about 35-40%, corresponds to wind vector displacements of about $0.8 \cdot 10^{-2} sec^{-1}$. When the wind vector displacements are greater than the stated limit, the probability of buffeting change only /120
slightly. Thus it has been established for the upper troposphere that the probability of buffeting is 35-40%, irrespective of intensity, when the vertical shear of the wind vector is $\geqslant 0.8 \cdot 10^{-2} sec^{-1}$.

High vertical wind vector shears occur in the lower stratosphere over the axis of tropospheric jet streams in the layer of sharp increase in wind velocity with altitude. During the winter, when the velocities of the west wind in the stratosphere are high, considerable vertical shears may occur in the zone of its stratospheric jet stream, the axis of which is located above the 50 mbar level in the middle latitudes. The critical values of vertical wind vector shear for cases of buffeting in the stratosphere remain uncertain at the present time. Analysis of known cases of buffeting at altitudes of 10-18 km [2, 3] indicated that the probability of moderate and

strong buffeting increases when the vertical wind vector shear is greater than $0.6 \cdot 10^{-2} sec^{-1}$.

The small scale structure of the wind fields have been investigated recently using special equipment. Such studies give a more complete presentation of the physical nature of turbulence. These measurements reveal the existence of a stable laminar structure of the wind field always to the flight altitude of supersonic aircraft [22].

The source of the zones of turbulence in the layers between 16 and 20 km (depending on the latitude) may also be the phenomenon of wind reversal (from west to east), which occurs in the summer in the middle latitudes of the stratosphere during transition from cyclonic circulation in the lower layers of the stratosphere to overlying anticylonic circulation (the level of wind reversal). In this layer of the stratosphere, which develops during the summer, there are usually weak winds of unstable directions. However, despite the low velocities, considerable vertical wind vector shears can develop due to sudden changes in the wind direction in the wind reversal layer, and this, in turn, can produce buffeting.

2. Temperature Distribution

The stratosphere is characterized, on the average, by considerably greater thermal stability than the troposphere. This is apparently one of the basic reasons leading to the reduction of recurrence of turbulence in the stratosphere. In individual cases, however, on the background of general stability of a rather thick layer of the atmosphere, turbulent zones may develop in comparatively thin layers, related to more or less substantial vertical temperature and wind gradients. In such layers density discontinuities may arise in the /121
presence of the corresponding wind shears, which often give rise to the turbulent condition of the atmosphere.

Radio sounding data in Sukhumi on the 14th of September, 1966 at 3 PM are presented in Figure 5.6 by way of example, and the zones of aircraft buffeting at various altitudes up to 18,000 m are indicated according to experimental flight data [2]. The flights were made over the Caucasus during the presence of discontinuity in the tropopause in this region. The zones of

aircraft buffeting above 11 km clearly corresponded to the levels of sharp discontinuities of the stratification curve in the presence of relatively vertical wind shears of about $0.6 \cdot 10^{-2} sec^{-1}$. Such thermal lamination with altitude was a result of advection of air masses of different origin: cold subtropical air at the 100 mbar level and above, and warmer air of the middle latitudes in the lower stratosphere.

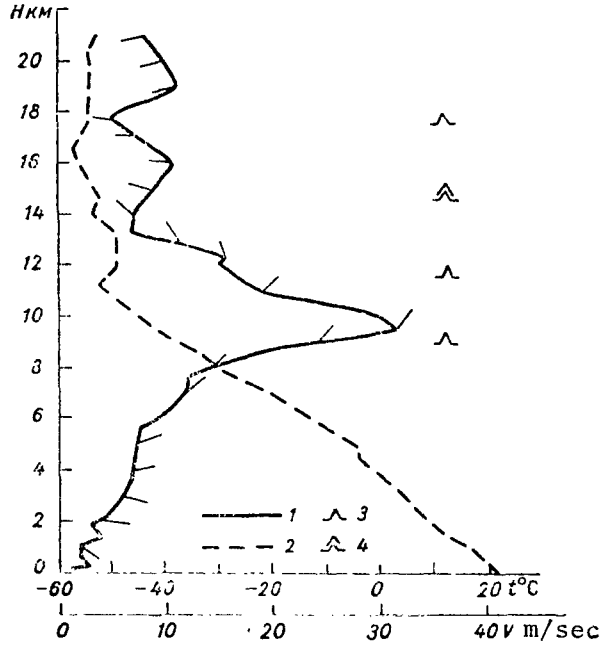


Figure 5.6. Profiles of Wind (1) and Temperature (2) According to Radio Sounding Data in Sukhumi at 3 PM of September 14, 1966, and Altitude Distribution of Moderate (3) and Strong (4) Aircraft Buffeting.

The analogous conclusions concerning the role of different thermal advection with altitude and the profile of thermal stratification in the formation of turbulence in the stratosphere were also given in [29, 34]. It was discovered during analysis of cases of turbulence in the stratosphere /122
over Australia that clear air turbulence is most common at the altitudes of sharp discontinuities of the temperature profile, which are the result of the different character of normal advection with altitude in the 200-50 mbar layer. The great importance of the vertical temperature distribution in the formation

of turbulence is apparently explained by the fact that sharp discontinuities of the temperature profile correspond to the location of baroclinic layers in the stratosphere, on whose boundaries in the presence of considerable vertical wind shear there exist conditions that are most favorable for the development of the turbulence [28]. In this connection the level of the isopause may be located in the middle latitudes of the zone of high turbulence recurrence [9].

The complexity of the thermal microstructure of the stratosphere is verified by special measurement data [7]. Measurements made with sensitive thermometers established that temperature gradients less than -0.8°/100 m and greater than +0.6°/100 m are encountered in 4-500 m thick layers of the stratosphere in less than 1% of the cases. Temperature gradients of ± 1°/100 m are encountered quite frequently in 100-150 m thick layers. Temperature gradients greater than 1°/100 m in absolute value are common in 20-30 m thick layers.

The cited data show that a combination of complex structures of the vertical temperature and wind distributions in the stratosphere can give rise to conditions that are favorable to turbulence.

Unfortunately, ordinary radiosonde measurements do not reveal completely enough the structure of the vertical temperature and wind distribution. The existing methods of smoothing off the temperature and wind profiles obscure the existence of comparatively thin layers of the atmosphere with sharp temperature and wind changes, and this makes it very difficult to analyze the conditions under which turbulence occurs.

Of great importance in the development of turbulence in the stratosphere, most likely, is the horizontal temperature field and in particular the magnitude, direction and level of reversal of the horizontal temperature gradient. The level of reversal of the horizontal temperature gradient is 10-14 km (first isopycnic) depending on the latitude and time of year [9]. Above this level the thermal wind blows in the direction opposite the baric wind, therefore the velocity of the latter decreases with altitude. The greater the horizontal temperature gradient above the isopycnic level the greater the negative wind shear. In the layer above the isopycnic level, in the HAFZ zone,

adaptation of the thermal and baric fields occurs and the stationary state of motions is interrupted, which can facilitate instability of the flow and formation turbulence.

The study of cases of aircraft buffeting in the stratosphere [21] shows /123 that turbulence in the HAFZ region is most common when horizontal temperature gradients are at least 2°/100 m.

According to published data [5], the probability of turbulence increases in the zone of tropopause discontinuity, which occurs, as we know, in the subtropics, and sometimes in the middle latitudes. Breakup of the tropopause presumably occurs only in the zone of intensive jet streams, which stimulate turbulence zones, usually in clear skies. The recurrence of clear air turbulence at altitudes of 12-20 km should be higher in the southern latitudes of the USSR, where the westerly subtropical jet stream, the axis of which is located at an altitude of about 14 km and where the zone of transition of the tropical tropopause to the tropopause of the middle latitudes exists year round.

3. Physical Model of Turbulence in the Stratosphere

It was stated above that buffeting is related to zones of increased vertical and horizontal temperature and wind gradients. It is obvious, however, that large wind shears and temperature changes alone cannot cause buffeting of an aircraft (with the exception of certain cases when great irregularities in the horizontal temperature field can produce the effect of buffeting of an SST [30]). They only facilitate the development of strong vertical movements which produce buffeting. Some authorities are of the opinion [5, 22, 32, 36] that short gravitational waves (Helmholtz waves) form in the stratosphere in the presence of uneven temperature and wind distribution, the destruction of which produces turbulence. Gravitational waves usually occur at the boundaries between air masses of different density (inversion, tropopause, frontal boundaries) in the presence of a vertical wind shear. Their length varies from a few dozen meters to a few kilometers.

Flight in the zone of graviational waves may be associated both with cyclic buffeting due to sign-variable vertical movements at the peaks and

valleys of the waves, and with buffeting resulting from turbulence formed by the destruction of these waves.

The theory of stability of graviational waves is examined in [6, 11, 12]. It follows from theoretical considerations that gravitational waves should lose stability and change into turbulent eddies if the wavelength λ becomes less than some critical value λ_{cr},

$$\lambda_{cr}=\frac{\pi}{g}(\Delta u)^2\frac{1}{\sqrt{\left(\frac{\Delta T}{T}\right)^2+\frac{\beta(\Delta u)^2}{g}}}, \qquad (5.4)$$

where $\Delta u = |u_2 - u_1|$; $T = T_2 - T_1$; where u_2, u_1 and T_2, T_1 are the wind velocities and temperatures in the over- and underlying layers of air in relation to the boundary level. It is kept in mind that $T_2 > T_1$: /124

$$\beta=\frac{d\ln\theta}{dz}=\frac{2(\gamma_a-\gamma)}{T_1+T_2},$$

where θ is the potential temperature.

It is obvious from equation (5.4) that the greater the wind drop and the smaller the temperature drop in the over- and underlying layers, the greater will be λ_{cr} (i.e., the broader will be the region of wave instability). An increase in the critical wavelength is also facilitated by reduction in the vertical thermal stability β of the atmosphere. It should be noted that λ_{cr} depends more on dynamic than on thermal factors.

It has also been established that the stability of gravitational waves of a given wavelength depends on their direction of propagation [6]. The greatest instability takes place for waves propagating in the direction of the vector difference between the wind velocity in the over- and underlying layers, while waves propagating perpendicular to this direction should be absolutely stable.

The mechanism of formation and destruction of gravitational waves that lead to the formation of turbulence of the various intensity can be established by means of meso- and microscale measurements of the wind and temperature fields in the various synoptic positions.

One of the causes of aircraft buffeting may be orographic waves of disturbances in the atmosphere. These waves are formed on the leeward side of the mountains at a certain distribution of wind direction and velocity and temperature with altitude. The length of these waves is usually a few kilometers but can reach 25-30 km, and sometimes even more.

The theory of the formation of leeward waves has been investigated by many researchers (Kochin, Dorodnitsyn, Lear, Kenney, Scorer, among others) and will not be discussed here. We will mention only certain features that have a direct relationship to SST operation.

There is no doubt at this time that leeward waves may spread not only through the entire thickness of the atmosphere, but also penetrate into the stratosphere [10]. Even small mountains and hills may produce considerable disturbances, which can penetrate to a great height.

By way of illustration of the propagation of wavelike orographic disturbances into the stratosphere we cite the results of measurements made in the USA during investigation of leeward waves aboard several aircraft at altitudes between 5,000 and 21,000 m [27]. The location of the isentropes on February 15, 1968 in the region of the Rocky Mountains of Colorado, indicating the wavelike structure of the flow all the way up to the altitude of the 50 mbar /125
level (approximately 21 km), is shown in Figure 5.7. The length of the basic wave increases with altitude, and the shorter waves fade near the tropopause. In the stratosphere, according to U-2 measurements, high-frequency vibrations reappear, but largely in the form of unstable turbulent eddies. Moderate and strong buffeting were experienced in the stratosphere on that day.

The intensity of rising and falling movements in the mountain waves can reach great magnitudes. Thus, the vertical velocity of 27 m/sec was recorded by means of gliders at an altitude of 12 km over the Sierra Nevada Mountains (USA), and up to 40 m/sec aboard an airplane. The following data can be quoted as usual vertical velocities in orographic waves [22]:

Altitude in km.....	4.6-6.7	6.7-7.6	> 10
Vertical velocity, m/sec.....	10	5	2

The amplitude of the waves reaches 200-500 m, but it generally falls between 100-200 m. As the distance from the mountainous obstacle becomes greater and the altitude increases, the amplitude of mountain waves and the intensity of rising movements in them decrease. A reliable indicator of the existence of leeward waves are high-altitude cumulo-lenticular clouds (Ac lent.), which form the crests of the waves and appear as stationary bands or banks, extending parallel to the mountain range. /126

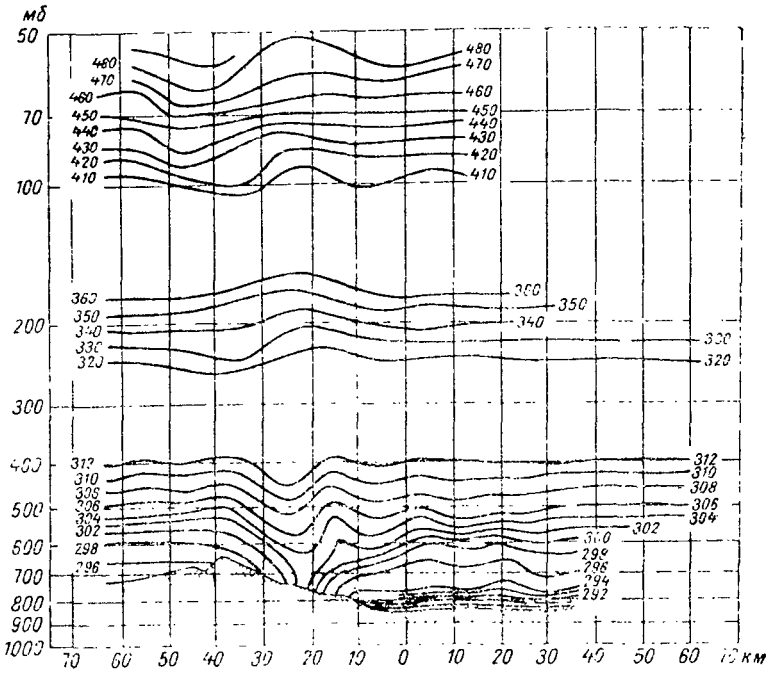


Figure 5.7. Potential Temperature Distribution with Altitude (in °Kelvin) on February 15, 1968 in the Region of the Rocky Mountains of Colorado According to Airborne Measurement Data.

Flight in such waves may be calmed with gentle rising and falling motions of the aircraft. Quite often, however, flight from the leeward side of the mountains is associated with moderate and strong buffeting. At high altitudes above the mountains the turbulence may be encountered as individually small zones that vary greatly in time and space.

Despite the great number of studies that have been conducted on turbulence over mountainous terrain, so far the basic causes of the destruction of mountain waves and the formation of turbulent eddies have not been established. Data of airborne measurements of buffeting over Australia are compared in [33] with the theory of leeward waves and turbulence. It is proposed that the energy stored in standing leeward waves may be transmitted in "cascade" fashion from waves of about 10 km in length to waves on the order of 100 km in length, which are also perceived as clear air turbulence. This physical model of turbulence, energized by mountain waves, permitted formulation of a prognostic scheme of turbulence over mountains. It should be pointed out that since orographic waves usually materialize in the presence of a certain distribution of wind and temperature with altitude and are localized on the leeward side of a mountain range, buffeting associated with them is easier to predict than buffeting caused by gravitational waves on the plane of a density discontinuity in the atmosphere.

It should be stressed once again that during analysis of baric topography maps of the upper troposphere and stratosphere over mountainous regions such as, for example, the Scandinavian mountains and the Rocky Mountains of North America, a vast zone of divergence of air streams is often observed. In such cases buffeting of an aircraft in the upper troposphere and stratosphere may be caused by turbulence that occurs during the action of mountain ranges on the air stream, and also due to turbulence in the divergence zone. Which one of these factors is most important is a question that probably depends on specific meteorological conditions and requires special investigation.

5.5. Forecasting of Aircraft Buffeting in the Stratosphere /127

Maps of the 300, 200, 100, 50 and 30 mbar baric topography, maximum wind maps and also data from individual radio sonde launches should be used for analysis and forecasting of aircraft buffeting in the stratosphere. The accuracy of baric topography maps (particularly forecast maps) and of data from individual radiosondes, is lower in the stratosphere than in the troposphere, and this must have some effect on the accuracy of diagnosis and forecasting of buffeting. The reliability of the forecasting of buffeting in

the stratosphere increases with the introduction to synoptic practice of objective analysis and numerical forecasting of the geopotential of isobaric planes of the upper levels and forecasting of temperature and wind for these levels.

It is also essential to use in actual practice maps of vertical wind shears, not only for the jet stream zones, but also for the entire territory.

So far, sufficiently reliable methods of forecasting aircraft buffeting in the stratosphere have not been developed, although analysis of cases of buffeting have made it possible to distinguish certain factors that promote turbulence in the stratosphere [2, 3, 21, 29]. Therefore only the most general recommendations can be made for diagnosing and forecasting buffeting in the stratosphere, based on previously examined causes of this phenomenon.

1. As already mentioned, during the winter in the middle latitudes of the stratosphere there are strong high-altitude frontal zones and associated jet streams. Usually they are of westerly and southwesterly direction. During this period the zones of turbulence will be related basically to the structure and configuration of the high-altitude frontal zones in which there are high vertical and horizontal temperature and wind gradients. Here attention must be focused on the presence of the higher values of horizontal wind shears along the stream in the zone of divergence and across the stream on the cyclonic side of the jet stream. The probability of buffeting increases at those levels where there is a considerable horizontal temperature gradient (> 2°/100 km), which often gives rise to substantial negative vertical wind shears.

2. In the summer, buffeting is most probable near the level of wind reversal (velopause) in the presence of high (> $0.6 \cdot 10^{-2} sec^{-1}$) vertical wind vector gradients.

3. When analyzing and forecasting buffeting in the stratosphere it is essential to analyze carefully the distribution of temperature and wind with altitude according to radiosonde data. Here the temperature and wind profiles should not be averaged and smoothed off. In the presence of high vertical wind vector gradients (> $0.8 \cdot 10^{-2} sec^{-1}$) the probability of CAT is much

higher. Conditions that are particularly favorable for buffeting exist near /128
thin intercepting layers and sharp discontinuities of the temperature stratification curve in the presence of the wind vector gradient with altitude.

4. The use of the Richardson criterion for detection of turbulence zones in the stratosphere, according to recent data, is possible [3, 5, 28, 29]. At this time, however, there is no universal opinion concerning the critical Ri numbers, below which turbulence can be expected in the stratosphere with a given probability. According to certain data [28] buffeting of aircraft in the stratosphere occurs when $Ri \leqslant 0.5$, but according to other data [29], when $Ri \leqslant 5$. According to data in [3] the possibility of buffeting is about the same for values $Ri \leqslant 10$, and drops sharply for $Ri > 10$.

It should be borne in mind that under conditions of great thermal stability of the stratosphere the Ri numbers should be determined basically by the vertical wind vector shear when calculating the Ri numbers on the basis of usual radio sounding data. In all probability the critical Richardson numbers are greater for the stratosphere than for the troposphere.

5. High recurrence of aircraft buffeting in the stratosphere should be expected in mountainous regions. The buffeting of aircraft in the stratosphere is most probable when the jet stream passes over a mountain range at some large angle to it, when the wind velocity at summit level is quite high ($>$ 10-12 m/sec) and when there is uneven vertical temperature distribution in the stratosphere, i.e., when conditions for the development of mountain waves exist. In this case buffeting may be encountered over mountains and on the leeward side of them.

In connection with the great difficulties in forecasting buffeting, both in the troposphere and in the stratosphere, special attention should be devoted to buffeting reports received from aircraft. With such data at his disposal the synoptician can determine quite accurately the zones of turbulence in the atmosphere and also make an inertial buffeting forecast, depending on how long the meteorological conditions continued under which buffeting was actually experienced.

To analyze and forecast buffeting in the stratosphere, and particularly over regions that are little known in the meteorological sense, it is advisable to use photographic data of the cloud cover obtained from artificial earth satellites. For example, in certain cases it is possible to locate (on the basis of satellite photographs) the jet streams and consequently the HAFZ in the upper troposphere, which can cover a considerable part of the stratosphere. Satellite photographs can clearly show the orographic cloud cover (Ac lent.) as parallel banks extending along the leeward side of the mountain range. The presence of such clouds indicates the development of leeward waves, which may be the cause of aircraft buffeting in the stratosphere.

Much hope has been placed recently, in connection with the difficulties of /129 forecasting for clear air turbulence, in the development of an aircraft instrument which would determine zones of turbulence in flight and warn the pilot of their approach.

5.6. Analysis of Individual Cases of Turbulence in the Stratosphere

We will examine certain cases of buffeting at high altitudes as example.

Example 1. September 3, 1966.

On that day a special scientific research flight was made over the Caucasus in the form of horizontal platforms at various altitudes between 9 and 18 km. The synoptic situation in the upper troposhpere and lower stratosphere was characterized by the presence of a high-altitude frontal zone, extending from the southwest to the northeast, and associated jet streams -- subtropical JS (jet stream) and JS of the middle latitudes (Figure 5.8). The axes of these JS in the flight zones were located not far from each other at different levels of the atmosphere.

During flight in the Sukhumi-Mineral'nyye Vody airlane the aircraft entered a zone of strong divergence of high-altitude streams with the branching of jet streams; here too there was buffeting in practically the entire layer between 9 and 14 km (Figure 5.9). The conclusions which I. G. Pchelko [15] derived concerning the importance of divergence of the streams in the formation of turbulence zones are thereby confirmed. The buffeting zone was also established at 18 km. The intensity of buffeting varied from moderate

to strong in all zones.

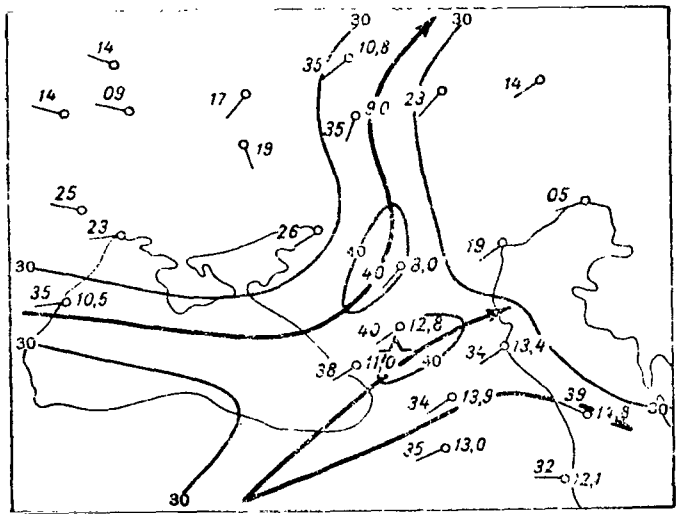


Figure 5.8. Map of Maximum Wind for 3 PM, September 3, 1966. Zone of moderate buffeting indicated in the region of Mineral'nyye Vody.

The magnitude of the horizontal wind shear along the stream on the 200 and 100 mbar isobaric levels was about 7 m/sec per 100 km. The temperature field at the 200 mbar level indicates that the horizontal temperature gradient in the region of the flights was 2°/100 km. There was a reversal of the horizontal temperature gradient between the 200 and 100 mbar isobaric levels. The vertical wind vector gradient in the same region was 0.5-2.0 m/sec per 100 m. /131

There was also a large vertical wind vector shift at an altitude of 18 km (2 m/sec per 100 m), basically due to a sudden change in wind direction. It is noteworthy that westerly circulation was reversed to easterly circulation at this altitude, i.e., the buffeting that was recorded here developed precisely at the lower boundary of the velopause. The fact that turbulence was observed during this flight, the leeward side of the range (in the region of Mineral'nyye Vody) is proof that, in addition to the above listed factors,

the orographic effect, i.e., formation of mountain waves on the leeward side of the Caucasus, located perpendicular to the stream, can play a definite role in the formation of turbulence.

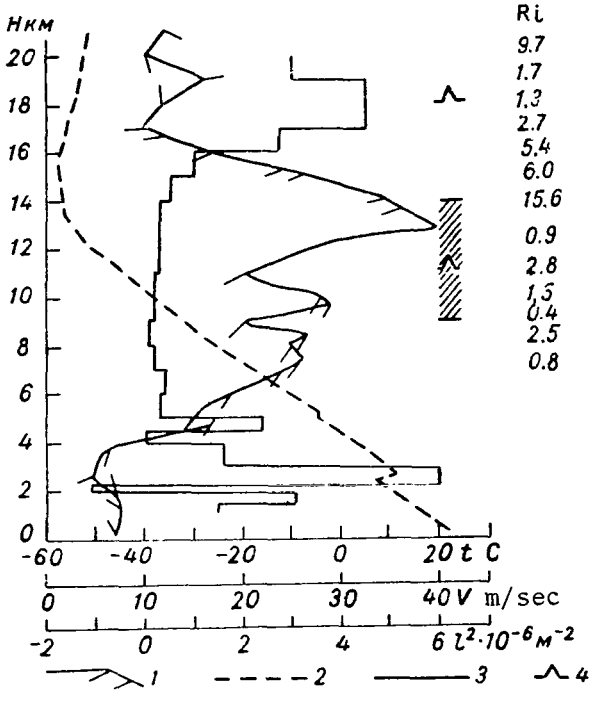


Figure 5.9. Profiles of Wind (1), Temperature (2), Parameter l^2 (3) and Distribution of Ri Numbers According to Radio Sounding Data at Mineral'nyye Vody at 3 PM, September 3, 1966, and Location of Zones of Moderate Buffeting (4). Wind direction indicated by arrows.

In order to prove this theory on the basis of radio sounding data in the Mineral'nyye Vody area the Scorer parameter l^2 was calculated (see Figure 5.9). According to Scorer's theoretical conclusions, atmospheric conditions are favorable for the formation of leeward waves if the parameter l^2 decreases with altitude; otherwise the waves should not form. If, however, the parameter l^2 does not change with altitude, then the theory does not answer the question of the possibility of formation of mountain waves. The l^2 distribution curve in Figure 5.9 shows that conditions for the formation of mountain waves existed in the lower troposphere at an altitude of 18-19 km (the parameter l^2 decreases with altitude). In the 9-14 km layer, where turbulence was observed, the parameter l^2 remain practically constant with altitude. However, judging from the fact that the wind velocities increased with altitude, reaching 40 m/sec, and the direction of the wind was perpendicular to the mountain range, it can be assumed that waves occurring in the lower troposphere could propagate into the higher layers of the troposhpere and the stratosphere.

The Ri numbers calculated on the basis of radio sounding data and plotted in Figure 5.9 have small values on the levels where aircraft buffeting was observed.

Example 2. April 18, 1967.

A flight was made on the Moscow-Ryazan airlane in the vicinity of a high-altitude frontal zone on the southern periphery of a stratospheric polar cyclone (Figure 5.10). At the 200 mbar level there was divergence of isohypses and isotherms with a reduction of wind velocity along the stream. The vertical cross section (Figure 5.11) shows that a frontal system, consisting of two fronts, corresponded to the high-altitude frontal zone in the troposphere. The maximum velocity of zonal flow occurred at altitudes of 8-9 km and it reached 60 m/sec in the area of the flight. The flight was made over the axis of the jet stream close to its cyclonic side. The tropopause was located at 9-10 km, dropping to the north of the flight region and rising toward the south.

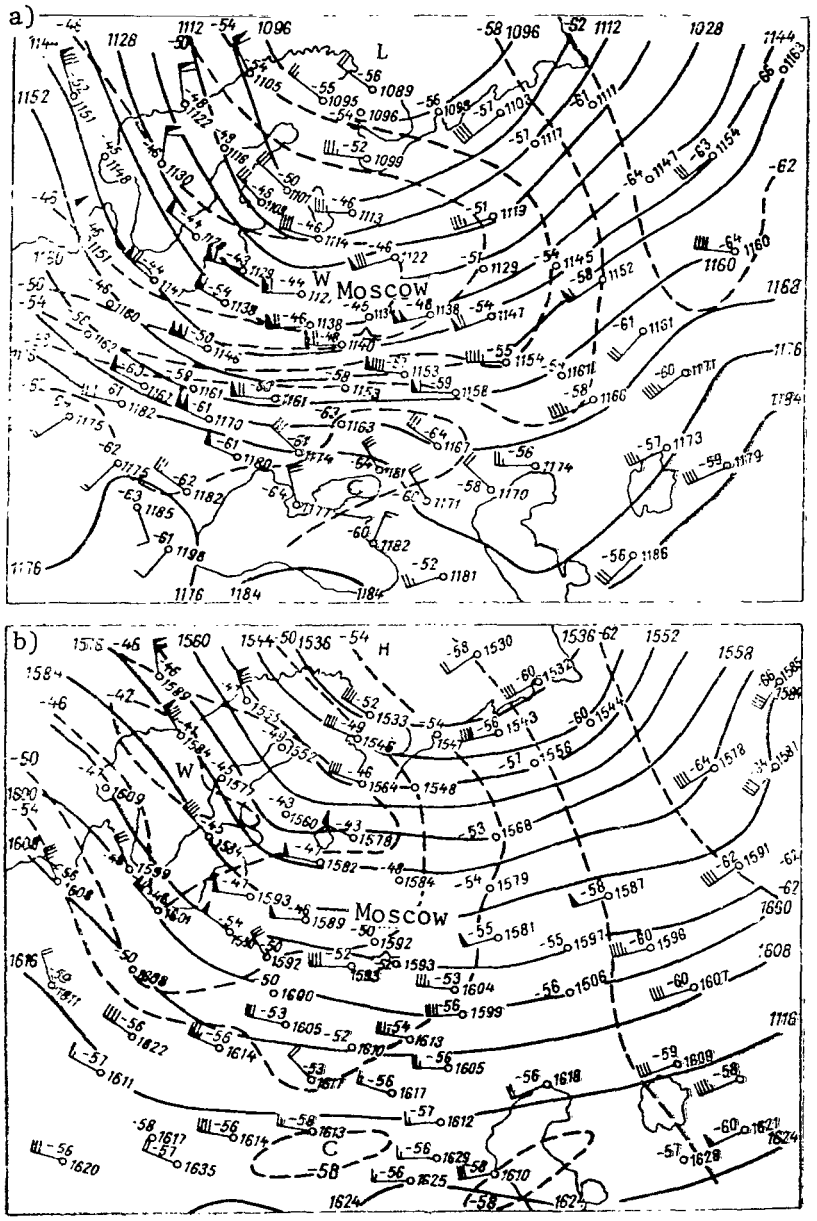


Figure 5.10. AT_{200} (a) and AT_{100} (b) Charts for 3 pm, April 18, 1967. Broken line denotes isotherms and zone of buffeting is indicated in the Moscow area.

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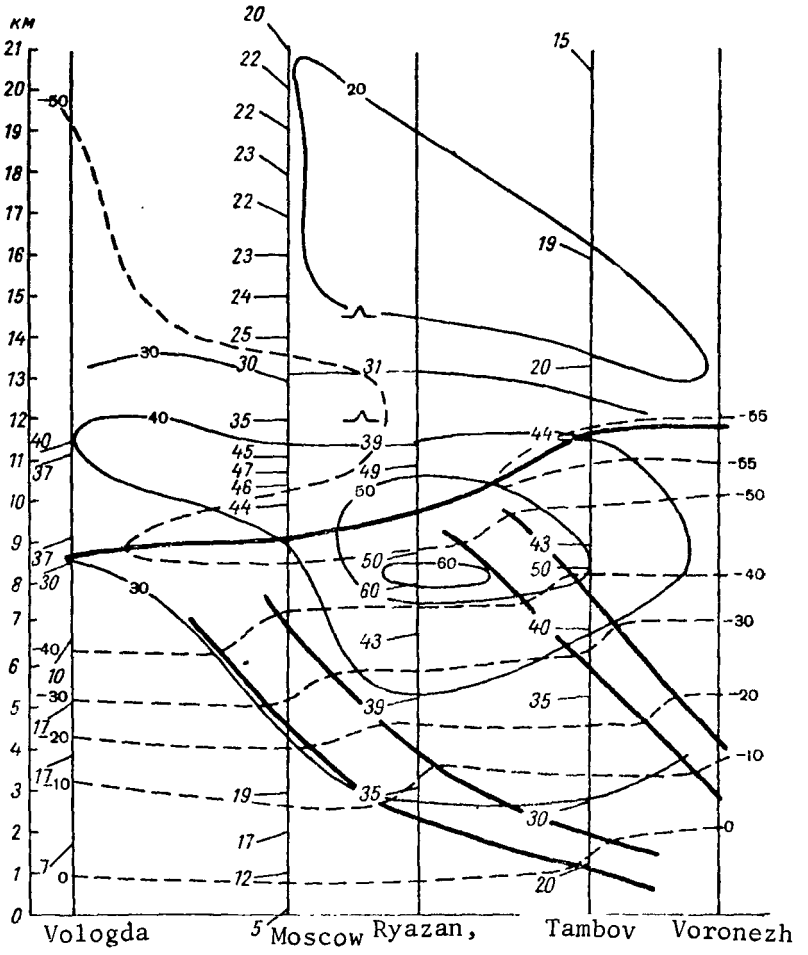

Figure 5.11. Vertical Cross Section of Atmosphere Perpendicular to Jet Stream at 3 PM, April 18, 1967.

The wind velocity in the lower stratosphere decreased rapidly and amounted to 22 m/sec at an altitude of 16 km. The zone of moderate turbulence, recorded /134 at an altitude of 11.8 km with short interruptions, extended for a distance of 65 km. The origin of this turbulence zone can be attributed to vertical wind vector gradients (up to 1 m/sec per 100 m) and also to the existence of a constant divergence of isohypses and isotherms in the 300-200 mbar layer.

A zone of moderate turbulence, 24 km long, was also recorded at an altitude of 14.6 km near the inversion layer in the stratosphere with a vertical wind shear of 0.5-0.6 m/sec per 100 m. There was no buffeting at altitudes of 16 and 17.6 km, where the vertical wind shear was close to 0.

We will discuss the horizontal distribution of the temperature field in the flight region. There was a horizontal temperature gradient reversal between the 300 and 200 mbar isobaric levels. At the 200 mbar level, on the cyclonic side of the jet stream near the axis of the trough, there was a warm region with a maximum temperature of -44°. To the south of this trough the air temperature dropped and reached -64° on the anticyclonic periphery of the jet stream. The horizontal temperature gradient in the lower stratosphere was directed from north to south at an altitude about 3°/100 km at the 200 mbar level. It gradually dropped with altitude to 1°/100 km at the 100 mbar level. Because of this temperature distribution the thermal wind in the lower stratosphere flowed opposite to the baric wind and the velocity of the actual wind at these levels dropped sharply, which was the cause of considerable negative vertical wind shear here. Thus, the greater the horizontal temperature gradient in the region of the high-altitude frontal zone the greater the vertical wind shear and consequently the greater the possibility of formation of turbulence and buffeting of aircraft.

Example 3. February 1, 1965.

On this day, at 2:35 PM, during a flight at an altitude of 17,000 m, the pilot reported strong buffeting 150 km southwest of Novosibirsk. The synoptic situation in the stratosphere in this case was similar to that examined in the preceding example. Buffeting ocurred in the vicinity of a high-altitude frontal zone that encircled the stratospheric polar cyclone,

with nearly straight isohypses, i.e., very little anticyclonic curvature (Figure 5.12). Considerable horizontal change in the wind field was not observed in the given case, and therefore it can be assumed that the basic cause of turbulence was vertical wind distribution.

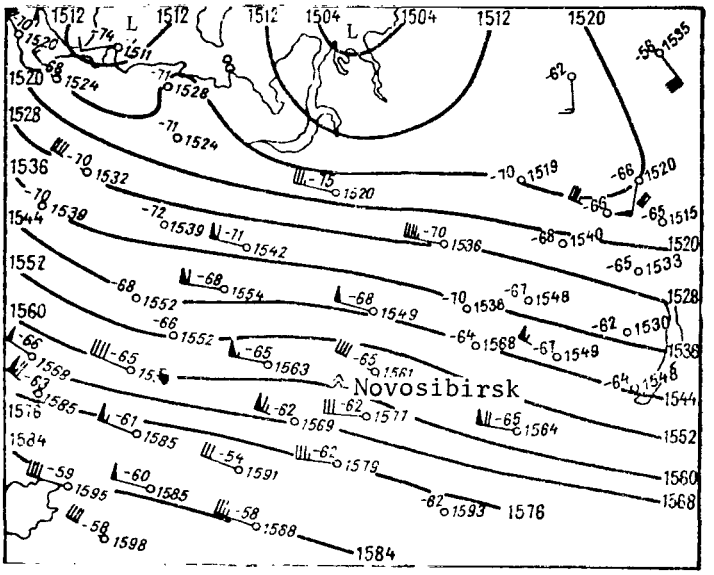

Figure 5.12. AT_{100} Chart for 3 pm, February 1, 1965. Strong buffeting noted northwest of Novosibirsk.

The horizontal temperature gradient retained its northerly direction with altitude in the stratosphere, and this facilitated the formation of a high-altitude frontal zone and strengthening of wind from 20 m/sec at the 100 mbar level to 40 m/sec at the 50 mbar level. The vertical wind vector /135 gradient in this layer changed from 0.6 to 1.1 m/sec per 100 m, and the Ri number was about 4. Also noteworthy in the given case was change in the character of temperature advection with altitude. If at the 200 mbar level there was advection of heat with isohypses of anticyclonic curvature, and 0 advection at the 100 mbar level, then advection of cold began at the 50 mbar level. The temperature dropped with altitude to -72°. The great importance of the change of character of temperature advection in the formation of turbulence was also noted above.

Example 4. April 20, 1969.

Extremely interesting data concerning the distribution of turbulence in the troposphere and stratosphere were obtained on April 20, 1969 during simultaneous flights of two aircraft equipped with scientific instruments: an IL-18 conducting research from 6,000 m-9,000 m, and a special high-altitude craft, flying between 10,000 and 18,000 m. The flights were made by the horizontal platform method with altitude increments of 1-2 km over the Caucasus and northeastern slopes toward Mineral'nyye Vody.

The IL-18, according to the authors observations, experienced maximum buffeting at an altitude of 7,500 m in clear sky. The strong buffeting, during which the aircraft drifted and lost altitude, began at a distance /136
of possibly 50 km from the mountain range on the leeward side and continued for 5 minutes during flight toward the range on a course of 210°. Near the range the buffeting weakened and almost entirely ceased. During flight near the range and parallel to it there were only individual disturbances, but moderate buffeting, intensifying toward the end of the pattern, was encountered while flying from the range on a heading of 35°. We point out that the As and Cs cloud cover on the leeward side took the form of individual broad banks parallel to the range. This indicates the presence of wavelike air movement. Buffeting, as a rule, intensified between the cloud banks, i.e., in the zone of downdrafts. Buffeting varied from weak to moderate at lesser altitudes, and principally in Cs-As clouds. Up to altitudes of 8,000 and 9,000 m there was no buffeting. During the second flight of the IL-18 on the same day buffeting was experienced in the 6-7 km layer.

Flight in the stratosphere was accompanied chiefly by weak and moderate buffeting on the leeward side of the range. At an altitude of 10,000 m buffeting did not occur, was weak at 12,000 m and increased at 13,000 m to moderate and at times strong (maximum overload increments reached ±0.5-0.6 g). At 14,000-16,000 m buffeting diminished and the overload increment did not exceed ±0.2 g, and at altitudes of 17,000-18,000 m there was no turbulence. As we see, turbulence in the troposphere and stratosphere is very irregular through altitude.

We will now examine the aerosynoptic conditions that existed on April 20, 1969, over the Caucasus. The examined region, located behind a cold front of moderate latitudes, which was rapidly moving from the northwest and passed over the Caucasus about 3 pm (Figure 5.13). The passage of the front was accompanied by multi-stage cloudiness and heavy precipitation. The lower boundary of the Ns was at an altitude of 600 m, and the upper edge of the Cs was located at 6,000 m of altitude.

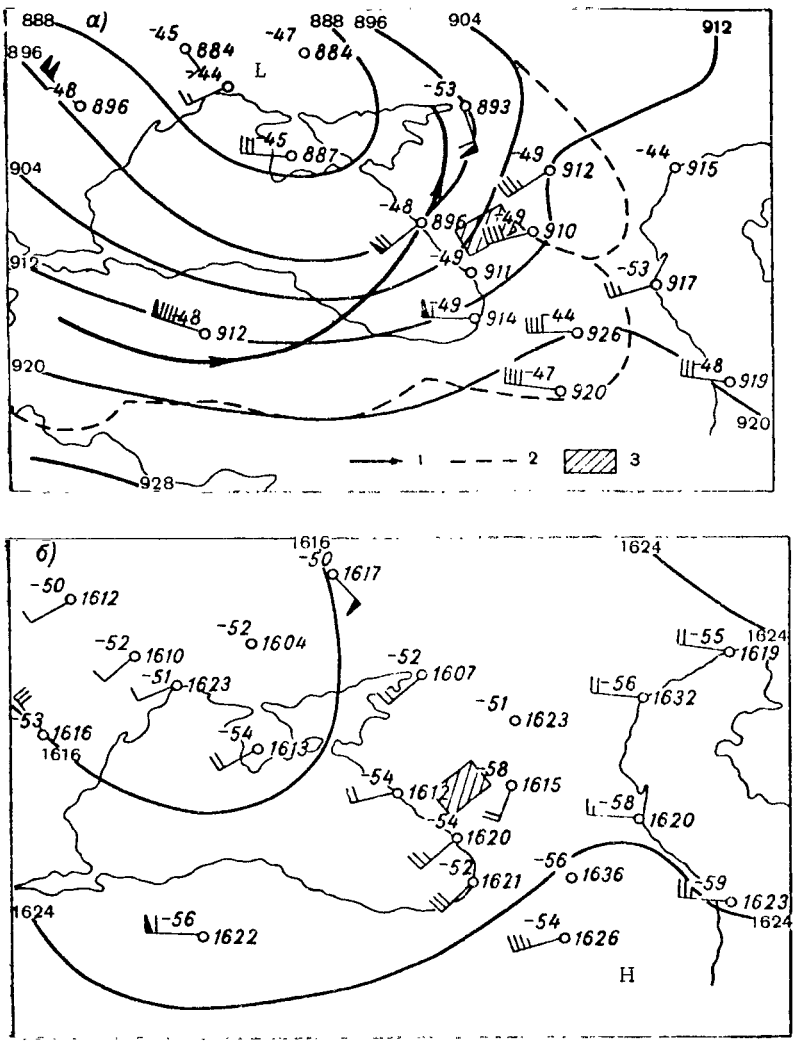

Figure 5.13. AT_{300} (a) and AT_{100} (b) Charts for 3 pm, April 20, 1969. 1 -- Jet stream axis, 2 -- location of front on the earth, 3 -- region of flights.

The delta of the high-altitude frontal zone, the axis of which extended from the southwest to the northeast, i.e., perpendicular to the Caucasus (see Figure 5.13), was located in the troposphere and lower stratosphere over the flight region. On the $OT^{500}_{1,000}$ (relative topography) map the leading part of the cold trough, moving from the northwest, corresponded to this delta. Because of this there was strong divergence of isohypses over the Causcasus and wind shear along the stream, which apparently, in combination with the orographic factor, was the basic cause of turbulence in the troposphere and stratosphere.

The vertical cross section (Figure 5.14) shows that the layer of strong buffeting in the troposphere was located at the upper boundary of the frontal plane beneath the maximum wind level. It should be noted that on the day of the flights there were favorable conditions for formation of mountain waves in the atmosphere. This came about because the wind at summit level of /139
the mountains was about 15 m/sec and strengthened with altitude, while its direction was perpendicular to the range. The isothermic layer was at an altitude of 7,000-7,500 m, a condition favorable to the formation of mountain waves, the existence of which was evidenced by the Ac lent. clouds observed during the flight. The altitude distribution of the Scorer parameter l^2 (see Figure 5.14) also indicates the possibility of the formation here of mountain waves (l^2 decreases with altitude). The Ri number in this layer drops sharply, and the vertical wind vector gradient increased. Thanks to the presence of a large vertical wind shear at this altitude (1.5 m/sec per 100 m) the wave motions could be destroyed and converted into turbulent eddies.

In the 12,000-14,000 m layer of the stratosphere there was also an increase in the vertical wind vector shear and a reduction of the Richardson number. Moderate and occasionally strong buffeting were encountered at an altitude of 13 km. At higher altitudes the vertical shear decreased, while the Ri number dropped sharply and turbulence abated. The horizontal wind shear in the delta of the high-altitude frontal zone in the lower stratosphere was 4-5 m/sec per 100 km. At an altitude of 13,500 m there was a sudden drop in the Scorer parameter l^2 which indicated conditons at this altitude favorable for the

formation of wave motions; moreover, as seen in the vertical cross section (see Figure 5.14), there was great temperature heterogeneity with altitude in the lower stratosphere, which also facilitates the formation of wave motions and turbulence.

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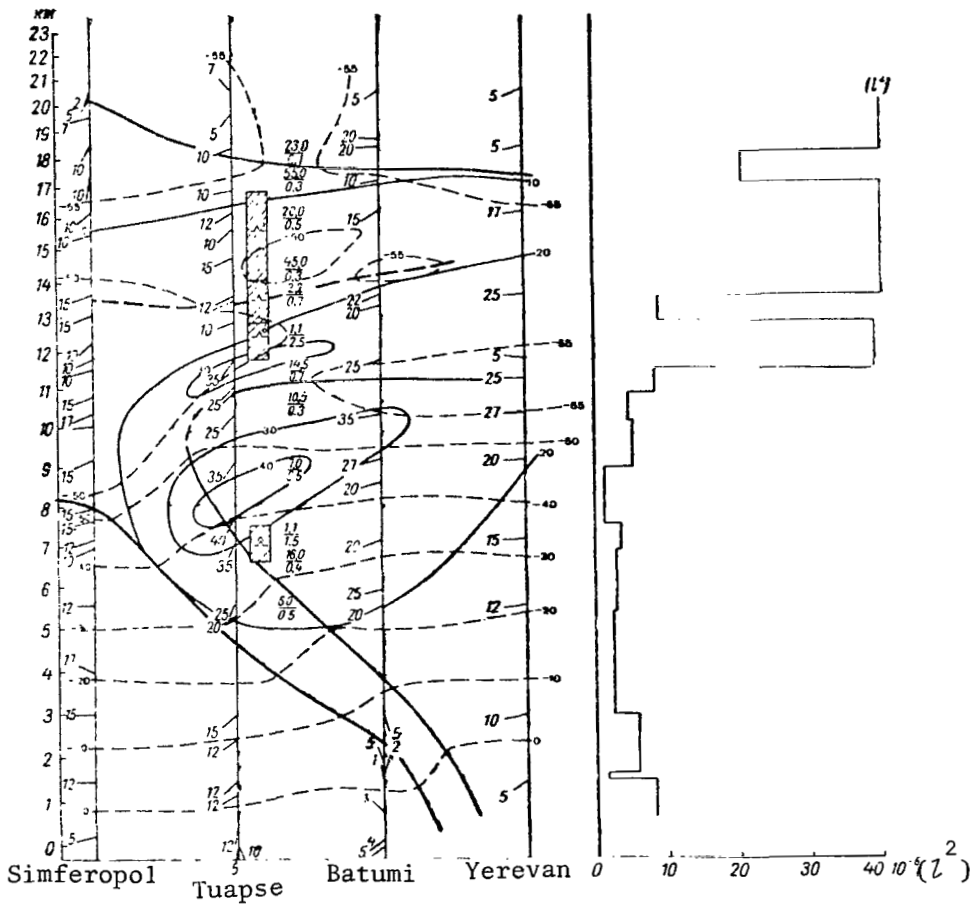

Figure 5.14. Vertical Cross Section Along the Simferopol'-Tuapse-Batumi-Yerevan Route at 3 pm, April 20, 1969.
The cross hatched rectangles indicate the distribution of buffeting zones by altitude. The number between Tuapse and Batumi give the Richardson numbers (numerator) and vertical wind vector shear (denominator). The altitude distribution of the Scorer parameter l^2 is given at the right-hand side of the figure.

CHAPTER 6 /140

EFFECT OF CONVECTIVE ACTIVITY, ICING, OZONE AND COSMIC RADIATION FROM SOLAR FLARES ON FLIGHT OF SUPERSONIC AIRCRAFT

6.1. Large Cumulonimbus Clouds, Thunderstorms and Hail in the Upper Troposphere

A supersonic aircraft flying in the near sonic phase should not approach large cumulonimbus cloud formations because of the danger of damage by lightning, also by impact with large raindrops and, particularly, hail.

The crest of large cumulonimbus cloud formations in the tropics, and sometimes in the subtropics, may reach the cruising level of the supersonic aircraft, particularly in the first half of the run.

There are several regions on the globe where the crest of large cumulonimbus cloud formations exceed 17-18 km during periods of rain and hail. These regions are southern Asia, the tropical equatorial parts of Africa, the Caribbean Sea and Gulf of Mexico, the Mississippi Valley in North America and the Amazon and Orinoco Valleys in South America.

Thus, in the Missippi Valley there are many cases of the penetration of the tops of cumulonimbus clouds into the stratosphere to an altitude exceeding 18 km during a thunderstorm [19]. Over western Bengal (India) on June 1, 1962 in the Calcutta area the maximum altitude of the upper boundary of cumulonimbus clouds was established by radar as about 21 km [16]. In the given case the development of heavy convective cloud cover occurred at the beginning of the summer monsoon season. In northern India, above New Delhi, the maximum altitudes of the crests of cumulonimbus clouds during the summer monsoon season of 1959 reached 17 km according to radar measurements [17]. The same altitudes of cumulonimbus clouds were observed visually over India /141
from a "Comet" subsonic jet aircraft in 1952-1953 [23]. On the Tashkent--Delhi airlane the upper boundary of cumulonimbus clouds during the summer (July-September) of 1960-1962, according to visual observations from a Tu-104 jet transport, sometimes reached altitudes of 14.5 to 15.3 km, even in the morning [15]. This means that it can be 1-2 km higher during the period of maximum convection.

Studies of damage to aircraft by hail in various regions of the world made it possible to establish the maximum altitudes at which large and small hail storms can be encountered. It was found that during the development of powerful convection in strong rising air currents hail of rather large dimensions can be encountered at all flight altitudes all the way up to the tropopause. Small hail stones are encountered in the crests of cumulonimbus clouds that penetrate 2,000 m or more above the tropopause in to the stratosphere. Thus it is possible for supersonic aircraft to encounter hail in the near sonic ascent phase at 9,000-12,000 m in the middle latitudes and at altitudes of 16,500-19,500 m in the tropics and while cruising.

As stated in Chapter 4, the poor maneuverability of supersonic aircraft at maximum flight speed does not permit avoidance of zones of large cumulonimbus clouds that are detected at close range. Therefore supersonic aircraft should be equipped with improved radar equipment, whereby zones of large convective clouds can be detected at great distances. Thus, for example, the Anglo-French supersonic Concorde aircraft will be equipped with a new radar which should be able to detect cumulonimbus clouds during flight to distances of 550 km. To prevent interruptions in observations this aircraft will be equipped with two radars. In the event of some breakdown in the system of the operating radar the backup system will be switched on automatically.

Supersonic flight near the crest of cumulonimbus clouds that penetrate into the lower part of the stratosphere is also undesirable due to the possibility of entering a zone of great temperature changes for a very short distance. For example, flight in clear sky next to strong centers of rain clouds in Oklahoma (USA) detected a temperature change of about 10° over a distance of approximately 2 km [19]. Such sudden temperature changes can be accompanied by a corresponding change in the Mach number, which can have an unfavorable effect on the operation of the jet engines and on the aerodynamic characteristics of the aircraft during flight at supersonic velocity at cruising altitude.

It follows from the above that, in addition to both airborne and surface radar observations, reliable forecasts of the vertical thickness of cumulonimbus clouds and of the probability of hail at the flight altitudes of supersonic aircraft are also essential. /142

6.2. Possibility of Icing

After take-off, a supersonic transport will gain altitude to 8-10.5 km and descend from this altitude prior to landing at a subsonic velocity of about 720 km/hour or more [18]. Therefore it is useful to analyze the results of analyses of the conditions of icing of present day subsonic aircraft, including jets, for preliminary evaluation of the possibility of icing of such aircraft in the ascent and descent zones at subsonic speed.

Icing of subsonic aircraft, according to research data [7, 10], occurs principally in supercooled liquid-droplet clouds and in precipitation at negative air temperatures from -3 to -12°C. Under these conditions the amount of ice that builds up on the surface of an aircraft per unit time, i.e., the intensity of icing, depends chiefly on the water content of the cloud and flight speed [9]. The humidity of clouds decreases as the temperature drops. When this happens the intensity of icing also decreases correspondingly. As flight speed is increased to 400-500 km/hour the intensity of icing increases at constant humidity. A further increase in flight speed causes a reverse effect: the increasing kinetic heating of the surface of an aircraft, the result of adiabatic compression of the surrounding air caused by it and friction with the passing flow, can lead to termination of icing and thawing. For the stated reason the recurrence of icing at speeds above 600 km/hour is 1/6 that at speeds up to 600 km/hour [7].

The kinetic heating of the leading surfaces of an aircraft at points of total stagnation of flow (where the velocity of air particles relative to the aircraft is 0) during flight in dry air at 800 km/hour, calculated by equation (4.4) is equal to 25°. At other points of the aircraft surface, where not all the kinetic energy of the flow is converted to thermal energy, heating will be less. This also pertains to the leading edge of a swept back wing, on which the relative flow of velocity does not equal 0. During flight in

supercooled clouds kinetic heating will be even less due to the evaporation of droplets that settle on the surface of the aircraft. Nevertheless, it remains high at a speed of 800 km/hour. This can be judged by the graph of icing as a function of the air temperature and aircraft speed (Figure 6.1) plotted on a basis of data from reports of crews of scheduled flights. This dependence has the form of a curve as plotted by the solid heavy line. At a flight speed of 800 km/hour icing occurred at a temperature of about -16°. /144
Consequently the actual kinetic heating of those parts of the aircraft surface where ice accumulates can be about 15°. It amounts to 0.6 of the kinetic heating at the points of complete flow stagnation during flight in "dry" air at a speed of 800 km/hour, calculated by equation (4.4). This relation apparently can be used for determination of the possibility of icing of a supersonic aircraft in the ascent mode prior to leveling off for cruise flight and in the descent mode prior to landing. The broken curve in Figure 6.1 makes it possible to determine the temperature of dry air (beyond clouds), at which the temperature at points of total flow stagnation at the leading surfaces of the aircraft at the corresponding flight speed is equal to 0°C.

If a supersonic aircraft gains altitude at an airspeed of 720 km/hour, then, according to the curve in Figure 6.1, icing can occur at an ambient air temperature not exceeding -12°. The recurrence of icing of subsonic aircrafts at -12° does not exceed 20-30% (Figure 6.2). As the temperature drops, it diminishes rapidly and does not exceed 10% at -20° [10].

Since the supersonic aircraft in the subsonic flight mode has tremendous /145
excess engine thrust, a slight increase in speed during ascent can prevent icing. For example, an increase in airspeed from 720 to 800 km/hour leads to termination of icing that has already started at an air temperature above -15°. If the air temperature is below this limit and a further increase in airspeed of aircraft at altitudes less than 8 km is not permissible, then it is possible to reduce the duration of icing in supercooled clouds by increasing the rate of ascent to the limit of human tolerance. According to ICAO data [18], the tremendous useful thrust of the engines of a supersonic aircraft permits ascent at a vertical rate of up to 35 m/sec (2,100 m a minute) at the beginning of ascent and up to 50 m/sec (3,000 m/min) at the end of the stage.

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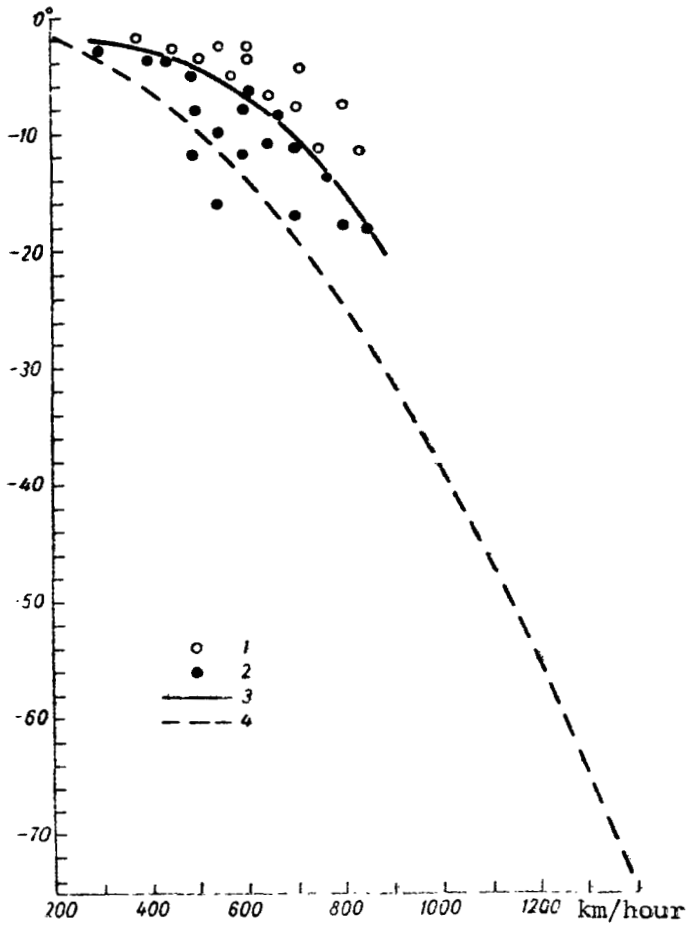

Figure 6.1. Icing of Aircraft as Function of Air Temperature and Flight Speed (According to I. G. Pchelko [7]).
1 -- Cases of no icing, 2 -- cases of beginning of icing of scheduled aircraft, 3, curve of icing, 4 -- curve of temperature of dry air corresponding to 0°C at points of flow stagnation on leading surfaces of aircraft.

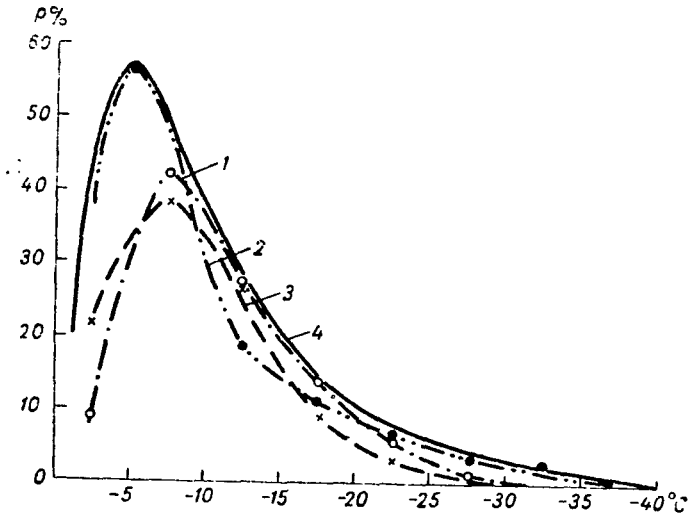

Figure 6.2. Recurrence of Icing as a Function of Air Temperature (According to O. Trunov [10]).
1 -- According to USSR Civil Aviation test flight data (1,775 cases), 2 -- according to questionnaire data from scheduled flights of BOAC aircraft and Tu-114 (260 cases), 3 -- according to foreign test flight data (220 cases), 4 -- envelope curve.

If supercooled clouds extend above 10.5 km then the possibility of icing becomes even less. The transition from subsonic to supersonic flight speed should occur in the 10.5-15 km layer. At the lower boundary of this layer the flight speed corresponds to Mach 0.93, at the upper boundary to Mach 1.6 [18]. At an airspeed of 950 km/hour icing is possible only at temperatures below -20°, and recurrence of icing at such a temperature does not exceed 10% (see Figure 6.2). At a flight speed of 1,350 km/hour a supersonic transport will not ice up, since the deposition of ice on its surfaces is possible only at temperatures below -39° (0.6 of the kinetic heating at the points of complete flow stagnation during flight in dry air is 39°) and the recurrence of icing at this temperature, according to Figure 6.2, is 0.

Thus the performance characteristics of the supersonic aircraft make it possible to reduce to a safe limit or to completely preclude icing in the ascent zone. At supersonic speeds of 1,400 km/hour and more this aircraft will be absolutely free of ice, since the temperatue of its surface will be positive

at air temperature in clouds up to -40° or below (depending on flight speed).

A supersonic transport may encounter icing chiefly in the descnet and landing zone. According to preliminary calculations an aircraft of the "Concorde" class will begin to descend from an altitude of 18.9 km 22 minutes prior to landing under normal conditions and 33 minutes prior to landing under conditions when the air temperature is 10° higher than in the standard atmosphere, and when there is a head wind of 25 m/sec [18]. At an altitude of 12 km (where icing is still possible) the aircraft will be under the influence of the stated conditions for 15 or 20 minutes prior to landing. If, for an even large part of this time, descent occurs in supercooled liquid-droplet clouds to an altitude of 100-200 m above the airport, then icing can begin only after cooling of the surface of the aircraft to -1°. When /146
the airspeed of an aircraft in the descent zone from altitude of less than 10.5 km is about 700 km/hour, this can occur in air with a temperature not exceeding -20° if only after several minutes. An increase in the air temperature and pressure with descent will prevent rapid cooling of the aircraft surface. Therefore it is difficult to expect that the duration of icing will exceed 5-10 minutes, and that the thickness of the ice on engine components and on the wings will exceed the permissible thickness for normal landing of a supersonic aircraft at a speed of about 250 km/hour.

In general the recurrence and the duration of icing of supersonic transports should be less than those of subsonic aircraft. The final conclusion, however, can be made only after special testing of such aircraft under flight conditions in supercooled clouds.

6.3. Ozone and Its Effect on the Aircraft, Crew and Passengers

At the earth's surface the air contains slight quantities of ozone (approximately $2 \cdot 10^{-8}$% by volume). Therefore, under normal atmospheric conditions we do not feel the strong oxidizing effect of ozone on the earth's surface. However, when the concentration of ozone in the air is high its damaging effect of certain materials used in aeronautical engineering, including rubber, and also on the respiratory organs of man and animal becomes

notable. It has been established experimentally that the maximum tolerable ozone concentration of man is $0.1 \cdot 10^{-6}$% by volume subjected to such conditions for 8 hours a day [14, 22].

It was pointed out in Chapter 2 that the maximum ozone concentration occurs in the stratosphere.

According to average data for the northern hemisphere, the ozone concentration in the atmosphere (Figure 6.3) at cruising altitudes of contemporary subsonic jet aircraft is several times greater than at their surface and approaches the maximum tolerable limit, and at cruising altitudes of supersonic transports exceeds this limit by 5 to 10 factors [22]. During the course of the year the maximum ozone concentration in the northern hemisphere occurs on the average spring in the high latitudes, and the minimum concentration in autumn in the middle latitudes.

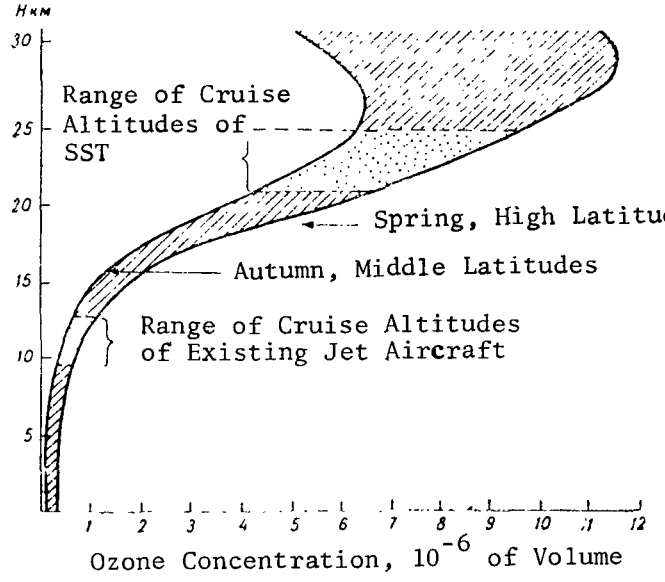

Figure 6.3. Average Vertical Distribution of Ozone in Northern Hemisphere (According to Dzhef and Istes, 1963) [22].

The vertical and horizontal distribution of ozone changes notably under the influence of the general circulation of the atmosphere. Here the

altitude of the layer of maximum ozone concentration also changes. Sometimes its lower boundary is several kilometers below the mean location.

Data from measurements of ozone concentration since 1957 sometimes indicate that the location of its maximum is lower than previously assumed. It is often found in the lower stratosphere directly above the tropopause [8]. An example /147 of high ozone concentration over Moscow below 20 km, according to measurements by the Central Aerological Observatory on April 17, 1958, is presented in Figure 6.4.

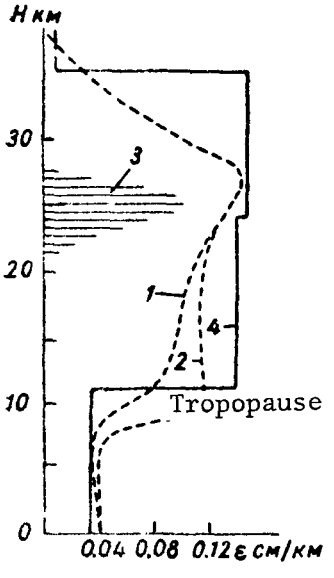

Figure 6.4. Vertical Distribution of Ozone and Nacreous Clouds (According to Driving) and Example of High Ozone Concentration Below 20 km According to TsAO Measurements During a Period of MGG (International Geophysical Year) [8]. 1 -- Vertical ozone distribution during days with substantial total ozone concentration; 2 -- during days with very high concentration; 3 -- recurrence of nacreous clouds; 4 -- ozone distribution over Moscow on April 17, 1958, according to measurement data of TsAO processed by A. P. Kuznetsov.

Other measurements made by the Central Aerological Observatory using special ozone probes in July, 1964 near the city of Ryl'sk, Kurskaya Oblast, show (Figure 6.5) how the vertical profile of ozone concentration in the atmosphere changes in connection with changes in the synoptic situation [2]. Thus, on July 12 (Figure 6.5 a) the observations were made during a period of increase in the overall ozone concentration (from 11-13 July) on the southern

periphery of a distintegrating anticyclone during reduction of the altitude of the tropopause and slight temperature rise in the stratosphere. On July 17 (Figure 6.5 b) the secondary maximum under the tropopause below the 8-11 km layer was observed toward the rear of the cyclone whose center was located over the northeastern part of European USSR, in addition to the basic maximum ozone concentration at an altitude of approximately 24 km. The formation of this maximum was possibly related to northwesterly flow of cold air. On June 19 (Figure 6.5 c) the vertical distribution of ozone in the eastern portion of the anticyclone, with a very high general ozone concentration, was characterized by a main maximum at an altitude of about 20 km, i.e., below the average location of maximum ozone concentration. On July 27 (Figure 6.5 d) /148 there were three maxima with approximately identical magnitudes at altitudes of 22, 16 and 6 km on the eastern periphery of a vast anticyclone over Western Europe.

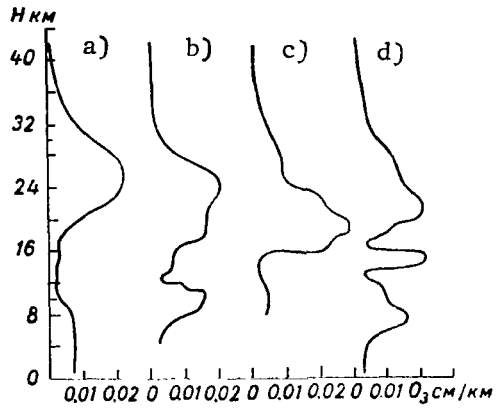

Figure 6.5. Vertical Ozone Distribution According to the Spectral Ozone Sounding Data, Ryl'sk, 1964 [2]. a -- July 12, b -- July 13, c -- July 19, d -- July 27.

According to data from US ozone probe observations, the layer of high ozone observation can usually expand downward in the high latitudes during winter and spring. In some cases it can descend to an altitude of about 5 km. Such was the case, for example, over Thule, Greenland (77° north latitude) on February 27, 1963 [14]. According to data from the same observations, the absolute maximum ozone concentration (by weight) reached $6.7 \cdot 10^{-6}$ at an altitude of 17.8 km on March 17, 1963 over Fort Churchill (59° north latitude).

It follows from the above examples and other material [1, 14, 22] that maximum ozone concentration will often occur in the middle and high latitudes near the cruise altitudes of supersonic transports. The probability of high ozone concentration in the lower latitudes is nonexistent, at least to an altitude of 17 km.

On the average the ozone concentration in the atmosphere at altitudes /149
between 18 and 20 km considerably exceeds $0.1 \cdot 10^{-6}$ parts per volume. However, before the air surrounding the airplane is heated inside the aircraft, it is compressed by a turbo-compressor to the normal pressure maintained in the passenger and service compartments of the aircraft. The air is heated as a result of compression. This destroys much of the ozone which it contains.

Experiments that have been conducted during flights of existing subsonic commercial jet aircraft have established that between 75 and 90% of the ozone in the air outside the aircraft, at cruising altitudes, dissociates before it enters the aircraft [22]. Other experiments have shown that it is not yet possible to establish whether or not there is a constant ratio between the ozone concentrations inside and outside the aircraft. Measurement data collected during a flight from New York to San Francisco on February 14, 1963 aboard an ordinary jet aircraft lucidly illustrates the close relationship between large fluctuations in ozone concentration as measured inside the aircraft, and its altitude relative to tropopause (Figure 6.6). At points where the tropopause extended below the cruise altitude, the ozone concentration was maximum and reached $6\text{-}9 \cdot 10^{-8}$ parts per volume in the given example. The ozone concentration in aircraft traveling the Anchorage (Alaska)-New York route

at an altitude of 11 km was 35-40·10^{-8} parts per volume in March 1963, i.e., three to four times greater than the tolerable level. During flights into the very same atmospheric conditions, the ozone concentration inside the aircraft could be much higher at an altitude of about 20 km.

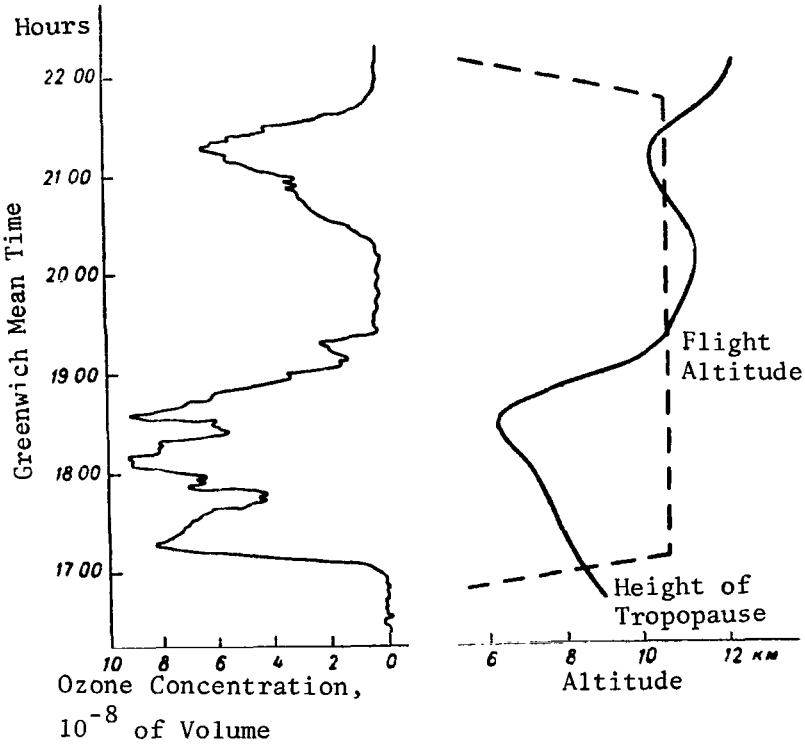

Figure 6.6. Ozone Concentration in Jet Aircraft During Flight from New York to San Francisco on February 14, 1963 (According to Brabets, 1963) [22].

In order to reduce the ozone concentration inside a supersonic transport to the established safe limit for crew and passenger, special filters must be used. If the use of such filters or other means of reducing ozone concentration inside aircraft will prove justified, then there is no need to provide current information concerning maximum stratospheric ozone concentration for supersonic flights.

Obviously, climatological data on the maximum ozone concentration near the cruise altitude during various months or seasons of the year along the principle airlanes are quite sufficient for evaluation of the effect of ozone

on the tires of supersonic landing gear. It is hard to expect that a high ozone concentration in the air flowing past the aircraft will have any great effect on the tires when they are tucked away in the fuselage in wings.

Kinetic heating of the air in the boundary layer of a supersonic aircraft should facilitate destruction of ozone, and thereby lessen its corrosive effect on the skin of the aircraft.

The ozone sounding network is still in the embryonic stage. Therefore vertical ozone concentration profiles are not available for all regions in /150
which supersonic aircraft may travel in the future. Meanwhile there is a regularly operating network of stations that measure the overall concentration of ozone. Studies have shown that an increase in the overall ozone concentration is a result chiefly of an increase in ozone concentration in the 12-24 km layer [1, 14, 21]. Consequently an increase in its overall ozone concentration in the atmosphere is, in the first approximation, an indication of an increase in ozone concentration at the cruise altitudes of supersonic transports. Maps of overall ozone concentration may be useful for this purpose.

Charts of the overall ozone concentration (in atmomillicentimeters) over the northern hemisphere, compiled daily by the USSR Hydrometeorological Center, make it possible to follow the movement of regions of relative to high or low ozone concentration [3]. The field of overall ozone concentration on the daily maps coincides with the temperature field on the isobaric planes from 50 to 10 mbar. At these levels, in the middle and high latitudes, regions of high ozone concen- /151
tration usually correspond to tongues of hate, and regions of lowest ozone concentration, to regions of cold.

Thus, on the basis of the temperature field at the 50 mbar isobaric level, it is possible to determine approximately where regions of high ozone concentration should be located.

Charts of mean monthly values of the overall ozone concentration for years (1964-1966) made it possible to uncover some characteristic features in its distribution. Thus, regions of high ozone concentration constantly prevail over North America. The ozone distribution during the course of the year is

less constant over Eurasia, where regions of high and low ozone concentration, distributed quite randomly, almost always alternate. The exception is maximum ozone concentration over Siberia. According to a small number of observations, low ozone concentration, compared to concentrations over the continents, prevail over the oceans.

The seasonal path of overall ozone concentration is characterized by the following features. During the cold months (November through March) there are two basic maximums: over Yakutsk and Canada. The centers of these maximums lie nearer the 60th parallel. By summer (April-July) these maximums move toward lower latitudes, particularly over Eurasia, and fade. Then a rather stable region of low ozone concentration materializes over the central Yakutsk plains and far east. It lasts through Autumn (October).

Charts of differences in the general ozone concentration for month to month indicate the general yearly trend in the ozone concentration. From December of 1964 to January 1965, for example, the increase in the overall ozone concentration was almost universally very pronounced. The ozone concentration continued to increase in the polar basin from March to April, whereas in the lower latitudes a stable decline had already clearly begun. A decline continued almost as universally through the subsequent spring and summer months. The chart of differences in overall ozone concentration between August and September indicates a stable region of increase in the north eastern part of the Asian continent. This region expands toward the polar basin from September to October, and from October to November regions of increase with great differences predominate. In December a nearly universal increase is noted.

Consequently it is easy to distinguish the principle features in the yearly trend for the year examined, 1965: the increase in overall ozone concentration from January to March-April, transition in early spring, and reduction, lasting until autumn of the entire hemisphere; during the autumn months an unstable situation develops, in which, along with vast regions of reduction, there are regions of increase; they begin to predominate in terms of area in November and especially in December.

Approximately 70% of all the ozone in the earth's atmosphere is located /152
between the tropopause and isopause [11]. The isopause is the boundary between the lower stratosphere with a slight vertical temperature gradient (isosphere) and the overlying layer of the stratosphere that is characterized by temperature inversion. The cause of the isopause is apparently the layer of maximum ozone concentration at altitudes of 20-30 km. The isopause rises as the layer of maximum ozone concentration moves upward and falls when this layer drops.

On the basis of a study made by the USSR Hydrometerological Center, the layer of maximum ozone concentration is located at the altitude of the isopause [11]. Therefore the altitude of the layer of maximum ozone concentration along the path of a supersonic aircraft and at moderate latitudes will be defined approximately according to the position of the isopause.

The position of the isopause is easy to find on the vertical temperature distribution curve, plotted on the basis of radio sonde data. Over the entire year the isopause is expressed chiefly through change in the vertical temperature gradient, and also through change in the wind direction and velocity in the summer. The temperature drops slightly in stratospheric cyclones between the tropopause and isopause, and it rises slowly in anti-cyclones. Above the isopause the temperature increases with altitude in stratospheric baric formations. During all seasons the isopause is lower in stratospheric cyclones than in anticyclones. As mentioned above, the layer of maximum ozone concentration is located at the same altitude where the isopause occurs. Therefore it can be assumed that the layer of maximum ozone concentration is higher in stratospheric anticyclones than in cyclones.

Seasonal changes in the isopause altitude are much greater in the high latitudes than in the middle latitudes. They are opposite to the seasonal changes of tropopause altitude and exceed the latter. The isopause is at its lowest position in January -- an average of 21.5 km in the cyclone and 25 km in the anticyclone. From January to April the level of the isopause rises from 21.5-25 km to 26-28 km and reaches its highest position in October -- an average of 30 km, both in the cyclone and anticyclone.

Additional information on conditions of ozone formation and distribution in the atmosphere are available in monographs and articles on this subject [4, 5, 12 and 13].

6.4. Solar Flares and Possibilities of Warning of the Cosmic Radiation Caused by Them

The dose of radiation produced by galactic cosmic rays and normal solar radiation is insignificant at an altitude of approximately 20 km and does not pose a threat to the crew and passengers of a supersonic aircraft at a /153
cruise altitude of 18-20 km [19, 20]. During the period of maximum solar activity, however, one of the large solar flares can produce a powerful flux of protons, possessing energies measured in tens, hundreds and even thousands of electron-volts. After a few hours this flux of high-energy protons reaches the upper layers of the earth's atmosphere. Their energy is sufficient, on collision with atoms and molecules of the upper atmosphere, to cause secondary high-energy particles, which can penetrate into the lower layers of the atmosphere and even reach the earth's surface. During a period of high solar activity a view of the strong flares, once or twice a year, can give rise to the analogous phenomenon, which will be recorded as an increase in the intensity of cosmic radiation. The intensity of such radiation, caused by solar proton flares, increases with altitude. At altitudes of 18-21 km a secondary high-energy protons can occur in numbers exceeding the tolerable limit of human safety, chiefly over the polar regions of the earth. Such dangerous doses of radiation may occur over these regions at an altitude of 24 km during the time of two or three of the largest flares during the entire 11 year cycle of solar activity. The intensity of the high radiation of proton flares weakens in the direction of the Equator, therefore the same dangerous level of radiation as occurs in the polar regions is possible in the middle latitudes only once or twice during the 11 year solar cycle.

The effects of the increase of intensity of solar flare radiation with altitude on a supersonic aircraft flying at Mach 2 poses less danger for the passengers and crew at an altitude of about 18,000 m than supersonic

flight at Mach 2.7 at an altitude of 21,000-22,000 m. Flight at an altitude of about 12,000 m is safe even during the largest solar flares.

It should be mentioned that for the passengers of a supersonic aircraft the probability of dangerous irradiation during two to three hours of flight is extremely slight. This risk is greater for the crew, since during the period of increased radiation, caused by one strong flare, the crew can make several flights; in addition, the crew of a supersonic aircraft may be subjected for a long period of time to irradiation by the less intense radiation of weak solar flares, which also eventually poses a threat to human health.

During the period of strong solar flares, however, it will be necessary to reduce the cruise level of supersonic aircraft to an altitude of 12,000 m. There is a need for rapid transmission to supersonic aircraft of information concerning the presence of solar flares and intensity of radiation caused by them at flight level. It has been proposed that solar flare forecasts be /154
used for preliminary planning of supersonic transport flights.

The radiation of proton flares is not a meteorological phenomenon. Measurement and prognosis of its activity are accomplished by special solar services. In the Soviet Union the solar service is a function of the Academy of Sciences. The solar service is also independent of meteorological services in other countries.

Recalling, however, the operational requirements of civil aviation, the 20th session of the executive committee of the World Meteorological Organization acknowledged that the telecommunications channels available to the meteorological services are probably most suitable for providing rapid transmission of warnings of solar flares, regardless of what agency originates them and where they are being transmitted. The desire to participate, in the event of necessity, in the promulgation of stated warnings for their use in aviation was expressed to WMO members.

The intensity of cosmic radiation at supersonic transport cruising altitudes increases to limits beyond the tolerable sanitary-hygienic dose of irradiation for people, often 15-30 minutes after the detection of a flare on the Sun. Therefore warning of the presence of a solar flare

and of a possible increase in cosmic radiation, received over meteorological communications channels, should be transmitted immediately by radio to the supersonic aircraft in flight as warning signals. After reception of such a signal the commander of a supersonic aircraft can lower the flight altitude to a safe level.

Since the intensity of cosmic radiation can remain high for several hours and even several days following the beginning of a solar flare, it is also proposed that the warning signals be transmitted to supersonic transport traffic control services.

Warning signals of the danger of proton radiation cannot always be received in time aboard an aircraft during flight, since strong radio interference follows soon after the solar flares, and sometimes long range radio signals on the daylight side of the earth are completely obliterated. The duration of total disappearance of radio signals is usually about 1 or 2 minutes. On November 19, 1949, however, not long after a flare was noted on the sun radio signals in the 15-60 m wavelength range vanished for almost 30 minutes [6]. Shortwave radio communications were disrupted over wide regions of the earth for 4 1/2 hours after the proton flare on the Sun that occured July 7, 1968.

Due to the fact that interruption of radio communications will prevent timely reception of warnings, it is proposed that an instrument be established /155 aboard supersonic aircraft for recording proton radiation, and which will signal the crew in the event that the permissible limit of dangerous radiation is exceeded [19]. The Anglo-French supersonic transport concorde is equipped with such an instrument. The data from this instrument will be useful for warning the crews of other supersonic aircraft in flight. The summarization of these data over a period of several years will yield reliable information concerning the maximum limits of proton radiation during solar flares at the cruising altitude of supersonic aircraft. They will also make it possible to refine the flight rules of supersonic aircraft and to improve the procedure of forecasting the danger of radiation from proton flares at altitudes of 18-24 km. At the present time scientists can judge the possible dosage of

proton irradiation at these altitudes only on the basis of approximate calculations according to data from the recordings of the energy of solar protons on artificial earth satellites and at the earth's surface.

In connection with forthcoming flights of supersonic transport planes in various regions of the earth it will be necessary to establish an operational international sun service. The realization of this measure, vital to international civil aviation, is obviously a goal of the coming years.

The International Civil Aviation Organization established a working group for investigation of the possibility of maintaining reliable radio communications between air traffic control services and aircraft in the airlanes through artificial earth satellites.

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